

CLIMATE CHANGE in DESERTS

Past, Present
and Future

MARTIN WILLIAMS

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CLIMATIC CHANGE IN DESERTS

Past, Present and Future

This book reconstructs climatic changes in deserts and their margins at a variety of scales in space and time. It draws on evidence from land and sea, including desert dunes, wind-blown dust, river and lake sediments, glacial moraines, plant and animal fossils, isotope geochemistry, speleothems, soils and prehistoric archaeology. The book summarises the Cenozoic evolution of the major deserts of the Americas, Eurasia, Africa and Australia and the causes of historic floods and droughts. The book then considers the causes and consequences of desertification and proposes four key conditions for achieving ecologically sustainable use of natural resources in arid and semi-arid areas. *Climatic Change in Deserts* is an invaluable reference for researchers and advanced students interested in the climate and geomorphology of deserts, including geographers, geologists, ecologists, archaeologists, soil scientists, hydrologists, climatologists and natural resource managers.

MARTIN WILLIAMS is Emeritus Professor at the University of Adelaide, Australia. His particular contributions to the field involve using evidence from a wide variety of disciplines to reconstruct prehistoric environments, ranging from the habitats occupied by early hominids in the Afar Rift of Ethiopia to the Neolithic occupation in the Sahara and the Nile Valley to the late Pleistocene wetlands in the arid Flinders Ranges of South Australia. He is a recipient of the Cuthbert Peek Medal from the Royal Geographical Society, the Sir Joseph Verco Medal from the Royal Society of South Australia, the Distinguished Geomorphologist Medal from the Australia and New Zealand Geomorphology Group, and the Farouk El Baz Award for Desert Research from the Geological Society of America. He is the author of more than two hundred research papers (twelve in *Nature*), and has edited and authored twelve books, including *Landform Evolution in Australasia* (with J.L. Davies, 1978), *The Sahara and the Nile* (with Hugues Faure, 1980) and *Quaternary Environments* (with David Dunkerley, Patrick De Deckker, Peter Kershaw and John Chappell, 1993, 1998).

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MARTIN WILLIAMS

The University of Adelaide



CAMBRIDGE
UNIVERSITY PRESS

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32 Avenue of the Americas, New York, NY 10013-2473, USA

Cambridge University Press is part of the University of Cambridge.

It furthers the University's mission by disseminating knowledge in the pursuit of education, learning and research at the highest international levels of excellence.

www.cambridge.org

Information on this title: www.cambridge.org/9781107016910

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First published 2014

Printed in the United States of America

A catalog record for this publication is available from the British Library.

Library of Congress Cataloging in Publication Data

Williams, M. A. J., author.

Climate change in deserts : past, present and future / Martin Williams, The University of Adelaide.
pages cm

Includes bibliographical references and index.

ISBN 978-1-107-01691-0 (hardback)

1. Paleoclimatology. 2. Arid regions climate. 3. Climatic changes – Environmental aspects.
4. Deserts – History. I. Title.

QC884.5.A73W55 2014

551.415–dc23 2014009755

ISBN 978-1-107-01691-0 Hardback

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Contents

<i>Figures</i>	<i>page vii</i>
<i>Tables</i>	xv
<i>Preface</i>	xvii
<i>Acknowledgements</i>	xix
1 Climatic change in deserts: An introduction	1
2 Present-day desert environments	13
3 Cenozoic evolution of deserts	22
4 Adaptations to life in deserts	37
5 Evolution of desert research	57
6 Dating desert landforms and sediments	77
7 Stable isotope analysis and trace element geochemistry	98
8 Desert dunes	112
9 Desert dust	142
10 Desert rivers	158
11 Desert lakes	189
12 The pluvial debate	209
13 Desert glaciations	224
14 Speleothems and tufas in arid areas	245
15 Desert soils, paleosols and duricrusts	258

16	Plant and animal fossils in deserts	283
17	Prehistoric occupation of deserts	302
18	African and Arabian deserts	328
19	Asian deserts	360
20	North American deserts	384
21	South American deserts	407
22	Australian deserts	423
23	Historic floods and droughts	449
24	Desertification: Causes, consequences and solutions	473
25	Current climatic trends and possible future changes	500
26	Towards sustainable use of deserts	516
	<i>Bibliography</i>	531
	<i>Index</i>	621

Figures

1.1.	Distribution of deserts and their semi-arid and dry subhumid margins.	page 2
1.2.	Dissected volcanic upland, Hoggar massif, central Sahara.	5
1.3.	Dissected sandstone uplands, Wadi Rum area, Jordan.	5
1.4.	Dissected sandstone plateau, or <i>mesa</i> , Arizona.	6
1.5.	Mount Connor, central Australia.	6
1.6.	Isolated sandstone hill, central Sahara.	7
1.7.	Uluru (Ayers Rock), central Australia.	7
1.8.	Granite inselbergs, Jebel Kassala, eastern Sudan.	8
2.1.	Schematic cross section of the global atmospheric circulation, showing location of the Hadley cells.	15
2.2.	Seasonal migration of global wind systems and of the Intertropical Convergence Zone (ITCZ).	16
3.1.	World tectonic plate boundaries.	25
3.2.	Cenozoic plate movements.	26
3.3.	Cenozoic sedimentation in the Atlantic, Pacific and Indian Oceans.	27
3.4.	Major tectonic and climatic events of the Cenozoic.	28
3.5.	The global hydrologic cycle (schematic).	29
3.6.	Orbital fluctuations.	33
3.7.	The last glacial-interglacial cycle (schematic). (A) is the last interglacial, (B) the Last Glacial Maximum and (C) the present interglacial.	34
4.1.	Afar women filling goatskin waterbags, Afar Desert, Ethiopia.	38
4.2.	Cross section through a <i>foggara</i> , central Sahara.	39
4.3.	Saguaro desert cactus (<i>Carnegiea gigantea</i>), near Tucson, Arizona.	41
4.4.	Influence of rainfall and soil texture on two acacia species, Sudan.	43
4.5.	Camels crossing sand dunes, south-central Sahara.	47

4.6.	Afar mother and daughter with edible lily bulbs collected from a waterhole in the Afar Desert, Ethiopia.	48
4.7.	Stick framework of a portable hut, Jubba Valley, Somalia.	49
4.8.	Completed hut, Jubba Valley, Somalia.	49
4.9.	Locusts on an acacia tree, Jebel Marra, western Sudan.	51
4.10.	Petra, the Nabatean city 'half as old as time', Jordan Desert.	52
5.1.	Two-thousand-year-old fragment of pottery tempered with the freshwater sponge <i>Eunapius nitens</i> (Penny and Racek), Jebel Tomat, lower White Nile Valley, central Sudan.	59
5.2.	Siliceous megascleres in sponge pottery, lower White Nile Valley.	60
5.3.	Gemmule membrane broken open to show interior, extracted from sponge pottery, Shabona, lower White Nile Valley, Sudan.	61
5.4.	Tamarix mounds indicative of shallow groundwater, Bir Sahara area, Western Desert of Egypt.	62
5.5.	Mud-brick fort abandoned as a result of climatic desiccation some 2,000 years ago in Xinjiang Province, north-west China.	68
6.1.	Radioactive decay curve showing exponential decrease through time in the relative concentration of a stable radioisotope, in this case radiocarbon (^{14}C) with a half-life of 5,730 years shown on the x -axis and 50 per cent shown on the y -axis.	80
6.2.	Geomagnetic time scale for the last 5 million years.	84
6.3.	Effect of contamination by modern carbon on radiocarbon age.	89
6.4.	Effect of contamination by inert carbon on radiocarbon age.	90
7.1.	Strontium isotopic composition of Blue and White Nile waters and of lakes in the White Nile headwaters.	105
8.1.	Sand dunes immediately east of the Aïr Mountains, south-central Sahara.	113
8.2.	Progressive evolution of a barchan dune into a linear dune, showing associated sand-moving wind direction.	113
8.3.	Map showing presently active desert dunes.	115
8.4.	Map showing desert dunes thought to have been active during the LGM.	116
8.5.	Wind velocity and sand movement.	118
8.6.	Source-bordering dune, lower Blue Nile, central Sudan.	120
8.7.	Model of a lunette formation as a lake dries out.	122
8.8.	Map showing active and fixed dunes in and beyond the Sahara.	125
8.9.	Map showing the dominant wind systems in the Arabian Peninsula and the location of the Rub al Khali and the Wahiba Sands.	127
8.10.	Map showing active and fixed dunes in the Thar Desert, India.	128

8.11. Stratigraphic section through a Quaternary polygenic dune in the Thar Desert, India, showing eleven alternating phases of soil/calcrete formation and sand accretion during the last 200,000 years.	130
8.12. Map showing sandy deserts (active dune fields: 1 to 7) and sandy lands (areas of stabilised dunes: A to E) in northern China.	131
8.13. Map of the late Cenozoic Kalahari Sands.	135
8.14. Map showing desert dunes, dune fields, lakes and rivers in the Australian arid zone.	138
9.1. Map showing modern dust source regions and major directions of dust transport.	145
9.2. Isolated sandstone hillock undercut by wind erosion near In Guezzam, central Sahara.	146
9.3. Map showing the global distribution of loess and desert dust.	150
9.4. Map showing the distribution of loess in China.	152
10.1. Pleistocene alluvium of the Awash River, southern Afar Rift, Ethiopia.	160
10.2. Alluvial terrace exposed on the flanks of a desert dune immediately east of the Air Mountains, south-central Sahara.	161
10.3. Ephemeral stream channel, Xinjiang Province, north-west China.	163
10.4. Ephemeral stream channel, Dire Dawa, southern Afar Desert.	163
10.5. Alluvial terrace, north of Aqaba, Jordan Desert.	164
10.6. Run-off and infiltration associated with deep and shallow soils.	165
10.7. Alluvial fan, Negev Desert, Israel.	170
10.8. Exposed side view of alluvial fan, Negev Desert, Israel.	170
10.9. Murrumbidgee paleochannels, Australia.	175
10.10. Blue Nile paleochannels, central Sudan.	179
10.11. Late Pleistocene meanders east of the White Nile, central Sudan.	180
10.12. The late Pleistocene Blue Nile.	182
10.13. The early Holocene Blue Nile.	182
11.1. Pliocene lake sediments (diatomites) in fault contact with older volcanic rocks, Afar Desert, Ethiopia.	191
11.2. Early Pliocene lake deposition in the Middle Awash Valley, southern Afar Rift.	192
11.3. The late Pleistocene Willandra Lakes in semi-arid New South Wales, Australia.	193
11.4. Pleistocene lake marls near Bir Sahara, Western Desert of Egypt.	199
11.5. Fluctuations in Lake Lisan, the Pleistocene precursor of the Dead Sea.	202
11.6. The Willandra Lakes viewed as a cascading system.	206

13.1. Mount Badda, Ethiopia, showing evidence of late Pleistocene glacial erosion.	226
13.2. Glacially eroded rock-basin lake with moraine dam at outlet, Blue Lake, Snowy Mountains, Australia.	230
13.3. Glacially striated bedrock, Snowy Mountains, Australia.	231
13.4. Erratic boulder transported by ice, Snowy Mountains, Australia.	232
14.1. Flowstone overlying cave breccia with Middle Stone Age fossils, Porc Epic Cave, Dire Dawa, Ethiopia.	246
14.2. Algal limestone pillars formed when Lake Abhe was full during the late Pleistocene and early Holocene, Afar Desert, Ethiopia.	255
15.1. Factors of soil formation.	260
15.2. Stone-layer formed by termite activity, Northern Territory, Australia.	269
15.3. <i>Nasutitermes triodiae</i> termite mounds, Northern Territory, Australia.	269
15.4. Fossil soils, Adrar Bous, south-central Sahara.	271
15.5. Paleosols associated with prehistoric artefacts and oil palm leaf fossils in the piedmont zone of Jebel Marra volcano, north-west Sudan.	272
15.6. Alluvial terrace composed in part of reworked loess, Matmata Hills, Tunisia.	275
15.7. Contorted Cenozoic salt lake sediments, Negev Desert, Israel.	281
16.1. Molar of Pliocene <i>Elephas recki</i> used to build stone enclosure for baby goats, Afar Desert, Ethiopia.	286
16.2. Pliocene pig mandible, Afar Desert, Ethiopia.	287
16.3. Surface shells on edge of Holocene Lake Boolaboolka, lower Darling Basin, Australia.	291
16.4. Changes in the proportions of aquatic, semi-aquatic and terrestrial snails in Gezira clay, lower Blue and White Nile valleys, central Sudan, between 15 and 5 ka.	292
16.5. Distribution of aquatic (left box) and land (right box) snail shells in Gezira clay, lower Blue and White Nile valleys, central Sudan.	293
17.1. The development of human culture during the late Pliocene and Quaternary, showing increasing 'hominization' through time.	303
17.2. Changes in hominid physical and cultural development from early Pliocene to late Pleistocene.	306
17.3. Origin and development of Palaeolithic stone tool technology from 2.5 Ma to 10 ka.	309
17.4. Short-horned Neolithic cow skeleton (<i>Bos brachyceros</i>), Adrar Bous, south-central Sahara.	312
17.5. Mesolithic and Neolithic stone tools from Adrar Bous, south-central Sahara.	318

17.6. Neolithic pots, Adrar Bous, Air Mountains, south-central Sahara.	319
18.1. Mean annual precipitation in Africa.	329
18.2. Surface winds and frontal locations (a) during July and August and (b) during December.	330
18.3. Saharan uplands.	332
18.4. Granite boulders exhumed from a deep weathering profile, Adrar Bous, south-central Sahara.	334
18.5. African and Arabian lithospheric plate movements and location of the East African Rift.	338
18.6. Main Ethiopian Rift, showing Quaternary lakes and volcanoes.	339
18.7. Dissected 30-million-year basalt flows near the headwaters of the Blue Nile, Semien Highlands, Ethiopia.	342
18.8. Generalised cross section across southern Africa.	355
19.1. Mean annual precipitation in Asia.	361
19.2. (a) Winter and (b) summer wind systems in Asia.	362
19.3. Cenozoic alluvial sediments tilted by recent tectonic activity, Xinjiang Province, north-west China.	365
19.4. River terraces, middle Son Valley, north-central India.	373
19.5. Cross-bedded late Pleistocene alluvial sands, middle Son Valley, north-central India.	374
20.1. Horizontal bedding of rocks exposed by erosion, Grand Canyon, Arizona.	385
20.2. Amphitheatre headwall created by differential erosion, Grand Canyon, Arizona.	386
20.3. Desert landscape, Arizona.	387
20.4. Major physiographic regions of North America.	388
20.5. Surface wind and pressure patterns (a) during the northern summer (July) and (b) during the northern winter (January).	390
21.1. Major physiographic regions of South America.	408
21.2. Mean annual precipitation, South America.	410
21.3. Surface wind, temperature and pressure patterns, South America (a) during the southern summer (January) and (b) during the southern winter (July).	411
21.4. San Pedro de Atacama and Licancabur, Chile.	420
21.5. Rio Loa, central Atacama, Chile.	420
21.6. Geoglyphs, northern Atacama, Chile.	421
22.1. Digital elevation model of Australia.	424
22.2. Major regions of Australia.	425
22.3. Present-day precipitation zones of Australia and surrounding region and major ocean currents.	426

22.4. The Indo-Pacific Warm Pool bounded by the 28°C isotherm.	427
22.5. Flinders Ranges, South Australia.	428
22.6. Kata Tjuta ('The Olgas'), central Australia.	429
22.7. 'Devil's Marbles' granite tor, Northern Territory, Australia.	429
22.8. Cenozoic environments of southern Australia.	432
22.9. Australia and New Guinea during the LGM.	444
22.10. Australia and New Guinea during the early Holocene.	445
23.1. Region influenced by the summer monsoon and the two key regions of the Southern Oscillation.	452
23.2. Time series representation of the Southern Oscillation Index (1870–1986), the annual rainfall index for China (1870–1979), rainfall over India in mm (1871–1985) and discharge in the Krishna River, India, in millions of MI (1901–1960).	454
23.3. Map showing the correlation between the annual rainfall index for China and the Southern Oscillation Index for June, July and August for the 1870–1979 period.	455
23.4. The ENSO index for the 1900–1980 period, with floods and droughts indicated for northern and central Thailand.	456
23.5. Correlation coefficients for the seasonal SOI against the tree ring index for <i>Tectona grandis</i> (teak) growing in Java, 1852–1929.	456
23.6. Nile River flood height at the Roda Gauge (1737–1903) showing correlation between droughts and/or years of narrow <i>Tectona grandis</i> (teak) tree rings in Java and years of below-average flow in the Nile.	458
23.7. Statistically significant correlations (significance shown in brackets below) between China rainfall, Indian droughts, Java tree rings, Nile flood height and El Niño occurrences in Peru for different time intervals between 1740 and 1984.	459
23.8. Time series (1941–2008) of average normalized April–October rainfall departure (σ) for twenty stations in the West African Sudan-Sahelian zone (11°–18°N) west of 10°E.	465
23.9. Hypothetical impact of overgrazing and reduced grazing, respectively, on plant cover, albedo and rainfall in drylands.	468
24.1. Linkages between global changes, human activities and desertification.	482
24.2. Examples of self-reinforcing mechanisms (positive feedbacks) at international, national and local levels resulting in rangeland desertification.	483
24.3. Abandoned homesteads on the Alashan Plateau, Inner Mongolia, northern China.	493
25.1. Solar radiation budget showing inputs and outputs.	502

25.2.	Absorption of solar radiation by certain atmospheric gases.	504
25.3.	Changing concentrations of carbon dioxide and methane from air bubbles trapped in Antarctic ice during the past 150,000 years.	506
26.1.	Reclaimed gully, middle Son Valley, north-central India.	524

Tables

1.1. Aridity zones defined by P/E_{pot} ratios.	<i>page 4</i>
1.2. Evidence used to reconstruct environmental change.	10
3.1. Cenozoic time scale.	23
3.2. Quaternary time scale (ka BP).	23
3.3. Global environmental changes of the past 130 million years.	24
3.4. Late Cenozoic tectonic and climatic events.	27
6.1. Dating methods commonly used in the reconstruction of climatic change in deserts.	78
10.1. Some attributes of fluvial systems in arid and in humid regions.	162
10.2. Erosion processes.	166
10.3. (a) The Blue Nile at 21–18 ka; (b) The Blue Nile at 15–14 ka.	183
15.1. Major international soil groups recognised in drylands.	261
15.2. Soil orders of North America.	262
16.1. Processes involved in the production of fossil pollen assemblages from parent plants and subsequent analysis and interpretation.	296
18.1. Late quaternary environments in the Sahara and Nile Basin.	348
22.1. Cenozoic vegetation and climate in Australia.	433
23.1. Data sets used in compilation of time series shown in Figure 23.2.	453
23.2. Statistically significant correlations between China rainfall, India droughts, Java tree rings, Nile flood height and El Niño occurrences in Peru for different time intervals between 1740 and 1984.	453
23.3. Documented occurrences of El Niño-Southern oscillation events in relation to droughts and fires in Indonesia (1877–1998).	457
23.4. Assumptions, testable conclusions and model results of the biogeophysical feedback models of the Sahel drought proposed by Charney (1975) and Charney et al. (1975, 1977).	467
24.1. Possible causes and consequences of desertification.	477

24.2.	Extent and severity of desertification in irrigated areas, rain-fed croplands and rangelands in the areas classed as drylands in Asia (in Thousands of Hectares).	489
24.3.	Extent of soil degradation in susceptible drylands (in Millions of Hectares).	497
25.1.	Summary of key greenhouse gases affected by human activities.	506
26.1.	Summary of global and regional environmental and climatic changes discussed in this volume from 250,000 years ago to present.	518
26.2.	Major soil groups in drylands and their susceptibility to various forms of land degradation.	523
26.3.	Major global environmental issues specified by Tolba and El-Kholy (1993).	528
26.4.	Primary aims decided at the Desert Margins Initiative, Nairobi, 23–26 January 1995.	529

Preface

The aridity, which makes life in deserts so difficult, has also preserved abundant evidence of a more humid past. For instance, scattered across the Sahara are numerous prehistoric occupation sites and rock paintings left behind by the Neolithic pastoralists who once roamed this now inhospitable land. Several thousand years older than the Neolithic art are the rock engravings and paintings of elephants, giraffes and other large herbivores that now inhabit the African savanna. Likewise, in the deserts that stretch from Arabia into Russia, China and India, deep rivers once flowed and freshwater lakes filled what are now dry salt pans. The same holds true for the Kalahari, the Atacama and the deserts of Australia and North America, prompting widespread curiosity about the climatic history of our deserts.

Apart from a natural concern over possible future changes in the earth's climate and their impact on the often poor communities of the semi-arid world, there are a number of reasons why a careful evaluation of what we currently know about the climatic history of the arid and semi-arid lands is both timely and worthwhile. One reason stems from plate tectonics, another from isotope geochemistry. An accurate long-term perspective on global climatic change has become possible as a result of recent advances in our understanding of world tectonic history. The combined evidence from deep-sea drilling, seismic investigations and paleomagnetic studies has allowed reconstruction of sea-floor spreading history and of continental apparent polar wandering curves. The data from land and sea are impressive and persuasive. The timing of late Cenozoic ice build-up in the two hemispheres is now known reasonably well, as are some of the associated changes in oceanic and atmospheric circulation, which are in turn related to the origin and expansion of the deserts.

Reconstruction of changes or fluctuations in oceanic circulation patterns used to depend very largely on sediment and microfossil studies. Analysis of the oxygen isotopic composition of the calcareous tests of suitable benthic and planktonic foraminifera now provides an additional and powerful means of assessing changes in ocean water temperature and salinity at depth and near the surface. After allowing for

local effects, it is also possible to use this technique to estimate changes in global ice volume. Times of lowest world temperature (glacial maxima) were times of greatest aridity in the tropical deserts and their margins, with massive export of desert dust offshore, even to central Antarctica.

Against this general background, this volume has three main aims. One is to examine critically the various lines of evidence from geology, biology and archaeology that have been used to reconstruct climatic change within the arid and semi-arid lands that now occupy some 36 per cent of the land area of the globe. If we include the dry subhumid regions of the world, since they, too, were once more arid than today, we are dealing with nearly 50 per cent of the land area of the Earth.

We also discuss Antarctica – the largest and driest of our cold deserts – and the Arctic, because they have long exerted a powerful influence on global climate. The second aim is to trace systematically the climatic history of the deserts from the inception of Cenozoic aridity some 30 million years ago through the fluctuations of the past 2.5 million years of Quaternary time until the droughts and floods of the present day. The final aim is inherently more speculative, but worthwhile withal, because it seeks to use the insights from our study of past events to envisage how human societies are likely to interact with possible future climatic changes in the desert world.

Acknowledgements

Over the years, many friends and colleagues across the globe have been generous with information and ideas about the desert world, often showing me first hand those starkly beautiful landscapes they have studied with such dedication. Dick Grove first suggested that I might like to replace him on an expedition to Jebel Arkenu in south-east Libya during the northern summer of 1962, an expedition that was organised by Captain (later Lieutenant-Colonel) David Hall RE from the Royal Military Academy, Sandhurst. Claudio Vita-Finzi gave me sage advice on desert travel based on his unequalled knowledge of the northern Libyan Desert.

A second expedition to south-east Libya the following summer, again led by David Hall, provided a welcome change from soil surveys along the Blue and White Nile rivers in Sudan, and enabled us to map and name two hitherto unexplored sandstone plateaux north-east of Tibesti volcano. The contrast between the harsh Saharan summer and the widespread evidence of a prehistoric human presence in that region from Early Stone Age times onwards led me to ask myself when, why and how often had the Sahara been a green and pleasant land and why it was no longer.

A move to Australia in late 1964 and a drive across the 'Red Centre' during a time of severe drought made me wonder about the causes of such droughts; the consequences were all too visible. One cattle owner told me that his eight-year-old son had never seen rain, reminding me of the road through Kufra Oasis, built of salt, in a land where rain fell once in fifty years. Later visits to the Thar Desert of north-western India, the Taklamakan Desert of western China, the Alashan Desert of Inner Mongolia, the Kalahari and Namib Deserts of southern Africa, the Afar Desert of Ethiopia, not to mention Arizona and the Grand Canyon, interspersed with recurrent visits to the drier parts of Ethiopia, Sudan, Niger and Kenya, and more sporadic visits to Algeria, Djibouti, Egypt, Somalia, Mauritania, Tunisia, Jordan and Israel, enabled me to continue my efforts to make sense of when and why our arid lands were once able to support more abundant life than they can today. Almost inevitably, such work led me to consider the causes and consequences of, and possible solutions

to, desertification processes, including the tantalizing question of how to distinguish between human impacts, whether direct or indirect, and ‘natural’ influences.

None of this research would have been possible without the support and encouragement of my companions and colleagues who shared my passion for desert landscapes. Many weeks of joint fieldwork with Desmond Clark in Niger, Sudan, Ethiopia and India; with Don Adamson in Australia, Sudan and Ethiopia; and with Mike Talbot in Niger and Australia forged ineffable memories. *Valete* Mike, Don and Desmond – you shared the joys as well as the hardships, and I remain the better for having known you.

In France, Théodore Monod, Jean Dresch, Hugues Faure, Françoise Gasse, Jean-Charles Fontes, Pierre Rognon, Georges Conrad, Maurice Taïeb, Nicole Petit-Maire, Michel Decobert, Jean-Louis Ballais, Mohamed Benazzouz, Jean Riser, Raymonde Bonnefille, Annie Vincens, David Williamson, Françoise Chalié, Pascal Lluch, Hélène Jousse, Hélène Roche, Edouard Bard, Xavier Le Pichon, Bruno Hamelin and many others offered wise counsel and spirited discussion, whether in the laboratory or in the field. In Germany, I benefitted from the impressive desert knowledge of Hans-Joachim Pachur, Stefan Kröpelin, Philip Hoelzmann, Baldur Gabriel, Horst Mensching, Dieter Jäkel and Helga Besler.

In India, Virendra Misra and S. N. Rajaguru inducted me into the subtleties of the Rajasthan desert landscapes, soils and prehistoric archaeology, following my earlier work in the Son and Belan valleys of north-central India with the late Professor G.R. Sharma and his team of keen archaeologists: V. D. Misra, B. B. Misra, D. Mandel, Jagannath Pal and Umesh Chattopadhyaya. The late Professor Liu Tungsheng guided my first footsteps in China, as did Yang Xiaoping, Wang Tao, Ci Long-Jun, Xiao Honglang and Li Aixin on subsequent visits. In the Russian Federation and beyond, I have benefitted from the wisdom and experience of Professors I. G. Gringof, G. S. Kust, I. S. Zonn and A. A. Velichko, and from the stimulation provided by many highly impressive younger researchers.

In Australia, many have contributed to my appreciation of this ‘flat brown land’, above all Joe Jennings, Jack Mabbutt, Bob Story, Bob Gunn, Bob Galloway, Jim Bowler, Rowland Twidale, Ron Paton, Bob Wasson, John Chappell, Brad Pillans, Peter Kershaw, Patrick De Deckker, Karl-Heinz Wyrwoll, Gerald Nanson, Paul Bishop, Paul Hesse, David Dunkerley, John Tibby, Cameron Barr, Vic Gostin, George Williams, Bob Bourman, Gavin Prideaux, Cliff Ollier and countless generations of students whose questions have kept me alert to new ideas and concepts.

My mentors, friends and colleagues in the United Kingdom have been a source of inspiration over the years: Dick Grove, Bruce Sparks, Dick Chorley, Andrew Warren, Claudio Vita-Finzi, Dave Thomas, Ian Reid, Jamie Woodward, Mark Macklin, Suzanne Leroy and Alayne Street-Perrott. The three months in 1970 working at Adrar Bous in the geographical heart of the Sahara with Desmond Clark, David Hall, Andy

Smith, Alan Pastron, Tony Pigott, Mike Saunders, John Rogers and Richard Trewby are engraved on my memory as one of life's formative experiences.

Visits to North America were enlivened by Tom Dunne, Stan Ambrose, Don Johnson, Dick Hay, Steve Porter, Alan Gillespie, Geof Spaulding, Minze Stuiver, Steve Warren, Estella Leopold, Bernard Hallet, Steve Burges, Kelin Whipple, Leal Mertes, Bob Balling, Will Graf, Ron Dorn, Vance Haynes and Vic Baker, all of whom were as hospitable as they were intellectually stimulating. As always, the field visits were greatly enlightening. Asher Schick, Ran Gerson, Yoav Avni, Aaron Yair, Hanan Ginat, Yehouda Enzel and Moti Stein have shared with me their unparalleled knowledge of the arid landscapes of Israel, and in Jordan during my 1975 visit to Wadi Rum and Petra – the 'rose-red city half as old as time' – the local Bedu spoke to me of Moses striking the rock to obtain water as if it were yesterday.

The people of the Sudan have invariably received me with generosity and courtesy during my many visits to 'the land between two Niles' and beyond. I owe particular thanks to Dr Abdelrazig Ahmed and Dr Yusif Elsamani (respectively, past and present Directors of the Geological Research Authority of the Sudan), Dr Yasin Abdl Salaam, formerly at the University of Khartoum, Professor Osman et Tom, former Director of Soils Research at Wad Medani, and Neil Munro, for their patience, wisdom and guidance during my more recent visits, since he who has once drunk the waters of the Nile must needs return.

I owe special thanks to Dr Emi Ito who provided informed advice, incisive comments and an invaluable exchange of publications during the early stages of writing; to Frances Williams for constructive criticism of each chapter and help with certain figures; to Christine Crothers for her meticulous drafting of the majority of the figures; and to Matt Lloyd at Cambridge University Press for editorial encouragement. Any errors, omissions and other solecisms remain my responsibility. The journey has been long but there was always something hidden beyond the ranges, and the golden road to Samarkand was ever there to entice the weary traveller.

1

Climatic change in deserts: An introduction

The desert shall rejoice, and blossom as the rose.
Isaiah 35.1

1.1 Introduction

In Book Four of *The Histories*, Herodotus (ca. 485–425 BC) repeats the tale of a group of people from the small town of Sirte in northern Libya who, goaded into an irrational fury by the south wind that had dried out their water storage tanks, declared war on the wind and marched into the desert, where ‘the wind blew and buried them in sand’. It was not always thus. A few thousand years earlier, numerous bands of cattle herders roamed what were then the vast grassy plains of the Sahara, and before then herds of African herbivores including antelopes, giraffes and even elephants had ventured into what was at that time a well-watered savanna landscape strewn with perennial rivers and freshwater lakes. The evidence of these past changes is still obvious to the observant traveller. Scattered across the 5,000 km width of the Sahara from Mauritania to the Red Sea is an abundance of beautifully executed rock paintings of Neolithic cattle, sheep and goats, as well as rock engravings of the wild herbivores, all of which were forced out by a progressively drier climate.

As the Saharan example shows, deserts are superb repositories of past climatic events. The very aridity to which they owe their existence has facilitated the preservation of landforms, sediments and soils developed under very different environmental conditions, as well as evidence of the former presence of plants, animals and pre-historic humans in areas now too arid to support much life. Contrary to the popular view of deserts as regions almost entirely covered by sand dunes – only a fifth of the Sahara is so covered – deserts are more likely to consist of rugged mountain ranges and dissected plateaux interspersed with vast gravel plains, intermittently active rivers and sporadically flooded lakes (Figures 1.1 to 1.6). Indeed, many of the landforms that are considered so characteristic of deserts are in fact inconsistent with present-day

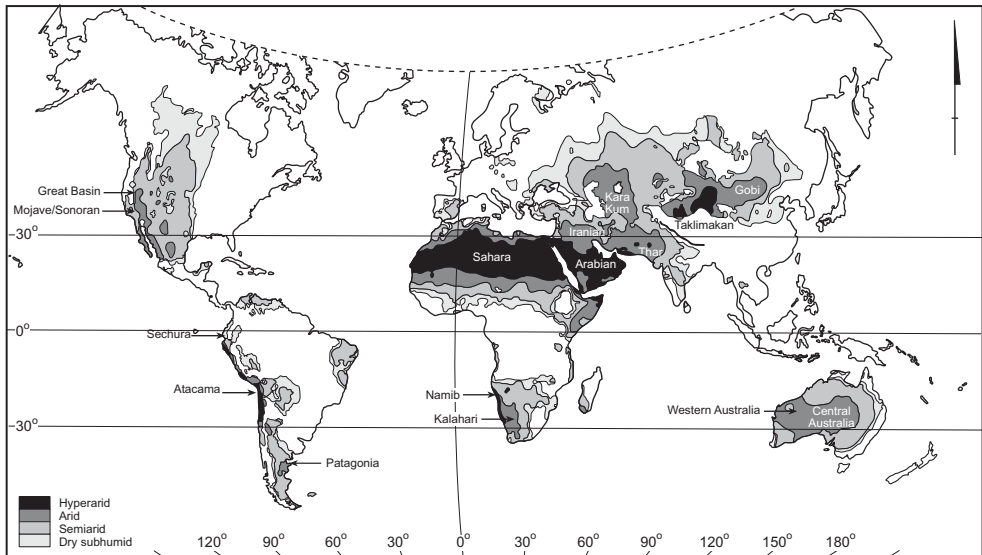


Figure 1.1. Distribution of deserts and their semi-arid and dry subhumid margins. (Adapted from UNEP, 1997, fig. 6.)

aridity, given that they are the results of weathering and erosion processes that are seldom active today. These observations invite us to ask when and why the deserts were once green and why are they no longer able to support much life. A further question is: How might they respond to future change?

Against this background, this volume has three main aims. One is to examine critically the various lines of evidence from geology, biology and archaeology that have been used to reconstruct climatic change within the hyper-arid, arid and semi-arid lands that presently occupy more than one-third of the land area of the globe. If we include the dry subhumid regions, that proportion increases to nearly one-half of the land area (see Figure 1.1). We also discuss both the Arctic – a region associated with globally important changes in ocean circulation initiated in the North Atlantic – and Antarctica – the largest and driest of our cold deserts – because Antarctica has long exerted a powerful influence on the global climate. The second aim, which follows logically from the first, is to trace systematically the climatic history of the deserts from the inception of *Cenozoic* aridity through the fluctuations of the *Quaternary* until the droughts and floods of the present day. (The *Cenozoic* covers the last 65 million years of geological time, with the final 2.6 million years being termed the *Quaternary*.) Our last aim is inherently more speculative, but nevertheless worthwhile, because it seeks to use the insights from our study of past events to envisage how human societies are likely to interact with possible future climatic changes in the desert world.

This introductory chapter enlarges on these aims, defines what is meant by a desert, outlines the approach adopted in this work, discusses briefly the scope and limitations

of the methods used to infer climatic change and introduces some of the key concepts analysed in later chapters. Our geographical focus is primarily on the tropical and temperate deserts and their margins, although, as we shall see, the cold desert of Antarctica has played a major role in the long-term desiccation of Australia (and possibly even central Asia). Antarctica has the distinction of being the coldest, driest continent on earth, with Australia being the second driest.

1.2 What is a desert?

Before proceeding further, it is worth considering what we mean by the term ‘desert’, starting with a very simple definition. For the purposes of this book, we can define a desert as a region where the precipitation is too little and too erratic and the evaporation is too high to allow many plants and animals to survive, except in a few favoured localities. Indeed, the Arabic word *sahra* denotes a flat wasteland devoid of water, to be traversed as quickly as possible. There is also an economic definition of a desert as a region where viable agriculture is not possible without irrigation – but this depends entirely on the type of crop being grown and begs the question of what is viable.

A more quantitative definition of aridity may be achieved using the ratio of precipitation (P) to evaporation. In practice, because long-term measurements of evaporation are rare for most deserts, evaporation is usually expressed as potential evaporation (E_{pot}). Potential evaporation may be calculated using the Penman (1948) formula, but here again there are too few reliable meteorological measurements to allow this approach to be widely used. The Thornthwaite (1948) formula is simpler to use and requires fewer climatic parameters. According to this formula, when $P = E_{\text{pot}}$ throughout the year, the index is 0. When $P = 0$, the index is -100 , and when P greatly exceeds E_{pot} , the index is >100 . Climates with an index below -40 are arid, -20 to -40 are semi-arid, and 0 to -20 are subhumid.

A somewhat arbitrary classification of aridity is that used by both the World Meteorological Organization and the United Nations Environment Programme, in which drylands are defined as those regions where the ratio of mean annual precipitation, P , to mean annual potential evaporation, E_{pot} , was less than 0.65 for the 1951–1980 period (UNEP, 1992a, 1992b). A modified version of the 1948 Thornthwaite formula was used to calculate P/E_{pot} . Using this approach, drylands are classed into hyper-arid, arid, semi-arid and dry subhumid, as shown in Table 1.1. Here again, we need to remember that mean precipitation is an almost meaningless concept in regions where the rainfall is so variable from year to year. It is also worth stressing that low precipitation is a *necessary* but not a *sufficient* cause of aridity. In certain cold areas of the world, such as Patagonia and Greenland, the rates of evaporation may be low enough to compensate for the low rates of precipitation, allowing a relatively dense plant cover and even peat bogs to exist in spite of a very low annual precipitation.

Table 1.1. *Aridity zones defined by P/E_{pot} Ratios*
(After UNEP, 1992a, UNEP, 1992b)

Climate zone	P/E_{pot} ratio	% of the world covered
Hyper-arid	<0.05	7.5
Arid	0.05–0.20	12.5
Semi-arid	0.21–0.50	17.5
Dry subhumid	0.51–0.65	9.9
Humid	>0.65	39.1
Cold	>0.65	13.5

In these instances, the *effective precipitation* is high enough to sustain plant growth, regardless of the absolute amount.

Table 1.1 shows that deserts and their semi-arid margins thus occupy 37.5 per cent of the land area of the globe, and if we include the dry subhumid regions, where mean annual rainfall may range from 750 to 1,500 mm, 47.4 per cent of the terrestrial surface. Given that roughly one in five persons now on this earth live in these drylands, it is important to understand how these lands have evolved through time and how they may change in the future.

1.3 Polygenic nature of desert landscapes

Desert landscapes are akin to ancient palimpsest maps in that they consist both of very young depositional landforms and of very old erosional landforms (Mabbutt, 1977; Frostick and Reid, 1987a; Cooke et al., 1993; Abrahams and Parsons, 1994; Thomas, 1997; Williams, 2002a; Laity, 2008; Parsons and Abrahams, 2009; Thomas, 2011; Goudie, 2013). The young landforms include dunes, alluvial fans, salt lakes and alluvial channels. The old landforms include mountains, hills and plateaux (Figures 1.2 to 1.6). It is misleading to assume that the landform assemblages that we find in present-day deserts has developed under entirely arid conditions. In fact, few have done so, because most of the major erosional landforms were shaped under previously wetter climates and have been preserved from further erosion by the onset of aridity.

Many desert landforms are exceedingly old. The vast desert plains of the central Sahara and western Australia have been exposed to subaerial denudation for far more than 500 million years, under very different climates from those prevalent today (Williams, 2009a). Desert monoliths such as Ayers Rock (Uluru) (Figure 1.7) in central Australia or the granite inselbergs of the Sahara (Figure 1.8), far from being diagnostic of aridity, owe their present morphology to prolonged and repeated phases of weathering and erosion under a succession of former climates, few of which were particularly arid. The abrupt juxtaposition of very ancient erosional landforms and



Figure 1.2. Dissected volcanic upland, Hoggar massif, central Sahara.

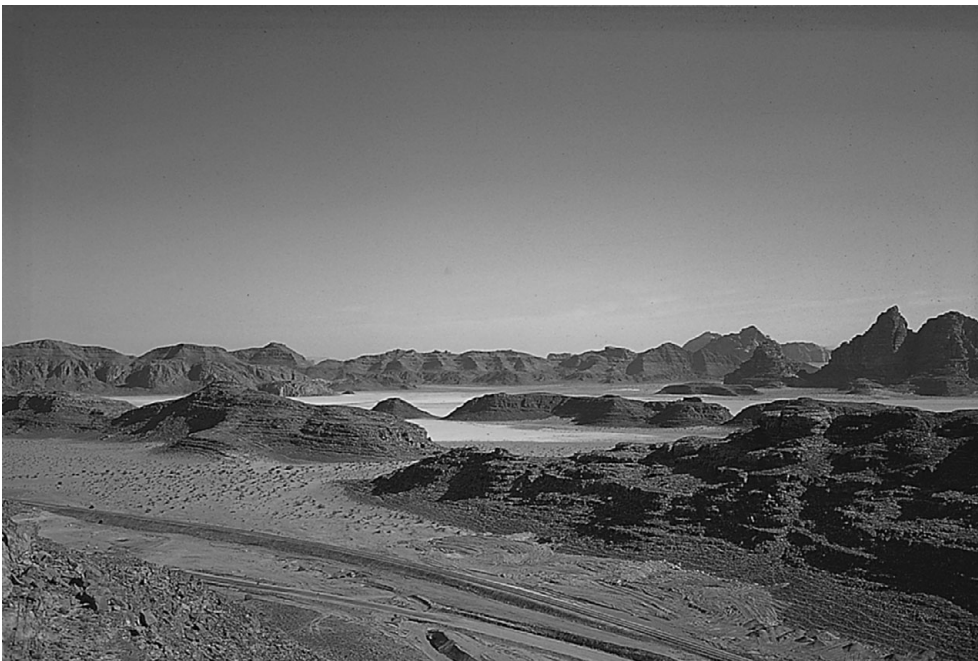


Figure 1.3. Dissected sandstone uplands, Wadi Rum area, Jordan.



Figure 1.4. Dissected sandstone plateau, or *mesa*, Arizona.



Figure 1.5. Mount Connor, central Australia.

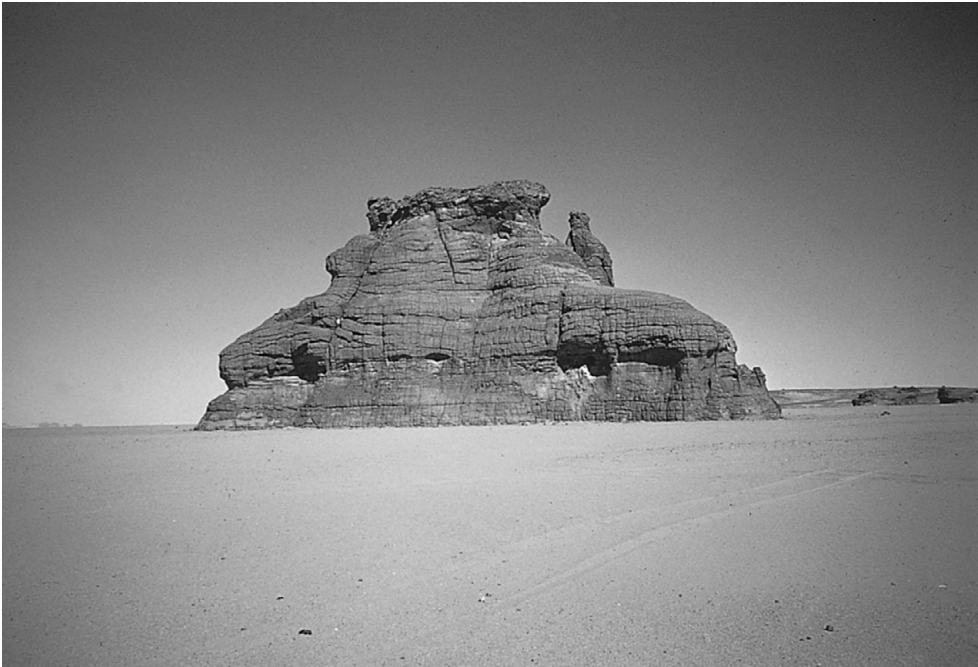


Figure 1.6. Isolated sandstone hill, central Sahara.



Figure 1.7. Uluru (Ayers Rock), central Australia.



Figure 1.8. Granite inselbergs, Jebel Kassala, eastern Sudan.

very young depositional landforms, together with the absence of vegetation and the sharp breaks of slope, give desert landscapes their peculiar and somewhat paradoxical character. These young landforms and sediments, whether eolian, fluvatile or lacustrine, contain the best record of past environmental changes, most notably the rapid climatic fluctuations of the late Cenozoic that provide the focus of much of this book.

1.4 Ambiguous quality of the evidence for climatic change in deserts

Early studies of desert regions tended to focus on specific desert landforms such as dunes, alluvial fans, river terraces, playa lakes and deflation hollows. In the last thirty years, particularly since the use of radiocarbon dating became widespread, paleoclimatic research in deserts has focussed on using alluvial and lacustrine deposits and their associated plant and animal fossils to reconstruct the history of desert rivers and lakes (Cooke et al., 1993; Abrahams and Parsons, 1994; Thomas, 1997; Parsons and Abrahams, 2009; Thomas, 2011; Goudie, 2013). One of the problems inherent in using high lake levels as evidence of formerly wetter climates lies in the complex hydrology of many desert lakes. Some are fed primarily from groundwater and may respond slowly to local changes in climate. Others may be fed solely from surface run-off. If the rivers that flow into these lakes originate in some distant, well-watered upland areas, the lake levels will fluctuate in response to distant changes in rainfall and may again not accurately reflect local conditions. Where the lakes are full and

overflowing and are merely enlarged portions of a through-flowing river system, they will also tend to be highly insensitive to local climatic fluctuations. Finally, is a lake high because of high rates of precipitation over the lake basin or because of much lower rates of evaporation related to colder or cloudier conditions?

Interpreting river sediments and landforms is equally fraught with ambiguity. Does widespread sedimentation reflect a river no longer competent to transport its load because of aridity in the headwaters and reduced discharge? Or does it reflect an increase in the supply of sediment from increased erosion in the headwaters, perhaps related to glacial and periglacial processes? Or might it represent a change from regular perennial flow to a more seasonal flow regime? To use a river terrace to infer a particular climate and then use the inferred climate to interpret other river terraces is to indulge in circular argument. None of these questions is easy to answer. Each requires accurate dating and careful scrutiny of many independent lines of evidence for its proper resolution. Throughout this work, we emphasise the different scales at which evidence of climatic change is to be considered, noting that the evidence is often fragmentary. The discerning reader needs to be fully aware of the scope and limitations inherent in the various proxies and archives used to reconstruct past changes in desert environments. For that reason, this book seeks to highlight the sometimes labyrinthine chain of reasoning involved in proceeding from environmental change to climatic change, noting that it is often more useful to know how the environment has fluctuated than to be overly concerned about distilling some imprecise climatic signal from inappropriate data. [Table 1.2](#) summarises the types of evidence used to reconstruct past environmental change and the variable of interest in this type of investigation.

There will be many cases in which a straightforward interpretation of past events is simply not possible with existing information. For example, it is perfectly feasible that quite different sets of processes can lead to the formation of a particular landform – a concept termed *equifinality* – so that the landform in question does not provide a clear signal as to how it formed. Likewise, a small initial perturbation can often trigger a *complex response*, one that is often unexpected. A simple example is strong wind scouring out a hollow in the lee of a small desert hill and eventually reaching the local groundwater table, so that a shallow lake comes into being without the need to invoke a wetter climate. This is easy enough to demonstrate experimentally but harder to show in the real world, because the groundwater table may have risen some unknown time after the deflation hollow was created. Another example of a complex response, again demonstrated experimentally, is the creation of a multiple set of alluvial terraces following the incision of a small channel under flume conditions (Schumm and Parker, 1973). Sounding a cautionary note to those of us involved in using river sediments to reconstruct Quaternary alluvial history, the authors of this elegant flume experiment found that ‘initial channel incision and terrace formation were followed by deposition of an alluvial fill, braiding and lateral erosion, and then, as the drainage system achieved stability, renewed incision followed by a low alluvial terrace’ (Schumm and Parker, 1973, p. 99).

Table 1.2. *Evidence used to reconstruct environmental change. (Adapted from Williams et al., 1998 and Williams, 2011.)*

Proxy data source	Variable measured
<i>Geology and geomorphology-continental</i>	
Relict soils	Soil types; isotopic composition of pedogenic carbonate concretions
Lakes and lake sediments	Lake level; varve thickness; facies changes; mineralogical composition; geochemistry
Eolian sediments: loess, desert dust, dunes, sand plains	Mineralogical composition; surface texture; geochemistry; provenance
Speleothems, tufas	Stable isotopic composition; geochemistry
<i>Geology and geomorphology-marine</i>	
Ocean sediments	Accumulation rates; fossil planktonic assemblages; isotopic composition of planktonic and benthic fossils
Continental dust; fluvial inputs	Mineralogical composition; surface texture; geochemistry; provenance
Biogenic dust: pollen, diatoms, phytoliths	Provenance; assemblage composition
Marine shorelines	Coastal features; reef growth
<i>Glaciology</i>	
Mountain glaciers; ice sheets	Terminal positions
Glacial deposits and features of glacial erosion	Equilibrium snow-line
Periglacial features	Distribution and age
Glacio-eustatic features	Shorelines
Layered ice-cores	Stable isotopic composition; physical properties (e.g., ice-fabric); trace element and microparticle concentrations
<i>Biology and biogeography-continental</i>	
Tree rings	Ring-width anomalies and density; isotopic composition
Fossil pollen and spores; plant macrofossils and microfossils;	Type; relative abundance and/or absolute concentrations; age; distribution
vertebrate fossils; invertebrate fossils:	Type; assemblage; abundance
mollusca; ostracods; diatoms; insects	Refugia: relict plant and animal populations
Modern population distributions	Phylogenetics; phylogeography
Molecular biology and genetics	
<i>Biology and biogeography-marine</i>	
Diatoms; foraminifera; coral reefs	Abundance; assemblage; trace element geochemistry; oxygen isotopic composition
<i>Archaeology</i>	
Written records; plant remains; animal remains, including hominids; rock art; hearths, dwellings, workshops; artefacts: bone, stone, wood, shell, leather	Age; distribution; morphology; provenance; geochemistry

1.5 Aims and structure of this volume

As noted at the beginning of this chapter, this book has three main aims, all of which are implicit in its title. The first aim requires us to consider very carefully the type of evidence used to reconstruct past climatic changes in deserts and desert margins. Until this is done, it is not possible to embark upon the second aim, which is to assess how deserts have responded to past climatic changes. The third aim is to consider how they might respond to future climatic changes. At this point it is pertinent to enquire just what the term *climatic change* denotes. The answer has to be a qualified one, given that it depends entirely on the scale (in time and space) at which the deserts are being studied. As we shall see, different climate proxies provide different levels of temporal and spatial detail, including the means of unravelling seasonal variations in precipitation.

Although deserts share a number of common attributes or diagnostic characteristics, each desert is unique and reflects the subtle interplay between local biophysical influences, including rock type, tectonic history, climate and biota. So as to avoid the pitfalls of over-facile generalisation, each of the world's larger deserts will provide examples to illustrate the discussion of past, present and future change.

In order to set the scene for what follows, [Chapter 2](#) deals with the causes of aridity, [Chapter 3](#) with the tectonic setting and geological evolution of the major deserts, [Chapter 4](#) with adaptations to life in deserts and with the influence of fire on the biota and [Chapter 5](#) with the history of climatic research in deserts. Because chronology is essential to any history of past climatic change in deserts, [Chapter 6](#) explains the need for careful dating and describes some of the more commonly used methods. Many attempts to reconstruct past environmental fluctuations on land and in the ocean rely heavily on isotopic analyses, so [Chapter 7](#) explains some of the isotopic and geochemical techniques most widely used in reconstructing past change.

The next nine chapters ([8](#) through [16](#)) consider the degree to which dunes, dust, rivers, lakes, glacial landforms, speleothems and tufas, soils, duricrusts and plant and animal fossils can provide precise and accurate information about past climatic change in deserts and desert margins. Desert margins are emphasised because deserts have been much more extensive at intervals in the relatively recent geological past, leaving behind a legacy of now fixed and vegetated sand dunes and other relics of their presence, such as mantles of desert dust or loess on which soils have since developed. One of these chapters ([Chapter 12](#)) summarises the reasons behind the persistent debate over the nature and significance of *pluvial* episodes in deserts. The use of archaeological evidence in interpreting past change and the role of prehistoric humans in modifying deserts environments are covered in [Chapter 17](#). The history of past climatic fluctuations in each of the major deserts is reviewed in [Chapters 18](#) through [22](#), starting with the impact of Cenozoic cooling and desiccation before proceeding to the millennial scale fluctuations of the past 2 million years and the

centennial to decadal fluctuations of the last 10,000 years. In these five chapters, we offer a synthesis of past environmental and climatic changes in the deserts of Africa (including the Arabian Peninsula), Asia, North and South America, and Australia. Topics covered separately in the earlier specialist chapters are brought together for the convenience of readers interested in those particular continents. Because the scope and limitations of the type of evidence used to reconstruct past changes have already been discussed in the topical chapters, the evidence is simply presented without further elaboration.

The pressing issues of historic droughts and floods discussed in [Chapter 23](#) lead us to consider the causes of, consequences of, and possible solutions to the thorny problem of contemporary desertification ([Chapter 24](#)). Indeed, the very word ‘desert’ comes from the Latin verb *deserere* (past participle *desertum*), meaning to abandon. Implicit in the term ‘desert’ is the notion that these now dry areas were once able to support more abundant life but have since become ‘deserted’. There is growing concern today that human actions are contributing to the spread of desert-like conditions in previously fertile and well-vegetated land, a complex set of processes known as ‘desertification’ (Mabbutt, 1978; Mabbutt, 1979; UNEP, 1992a; UNEP, 1992b; Mainguet, 1994; Williams and Balling, 1996; Williams, 2000; Williams 2002b; Williams, 2004). The causes of desertification are complex and often controversial; they include droughts, human mismanagement and the aftermath of war, and they are reviewed in the last three chapters of this volume. Two thousand years ago, the Roman historian Tacitus (ca. 56–ca. 120 AD) wrote scornfully of the scorched-earth policies favoured by some of the Roman emperors and their generals: *Ubi solitudinem faciunt pacem appellant* (‘They create a desert and call it peace’).

The Intergovernmental Panel on Climate Change (IPCC) volumes and cognate research papers have shown the nexus between ecosystem responses to possible future changes that could be triggered by global warming and the impact on societies living in the arid, semi-arid and dry subhumid regions of the world. [Chapter 25](#) reviews these issues, and the concluding [Chapter 26](#) provides a succinct but robust set of guiding principles for achieving ecologically sustainable use of deserts. Because nearly half of the land area of the globe is considered dry or seasonally dry (see [Table 1.1](#)) and provides a home for about a fifth of the world’s present human population, these are not trivial matters.

Whether reconstructions of past climates can provide us with a useful template for assessing likely future changes is hard to gauge, but given that global climate models are of necessity limited by the assumptions upon which they are based (which may or may not be valid), it would be unwise to ignore what has already occurred. Indeed, one could argue that knowledge of past events is often our only reasonably sure guide to what might happen in the future.

2

Present-day desert environments

El sharia el howa

The way of the wind

Arabic expression for desert tracks known only to smugglers and locals

2.1 Introduction

Before considering why deserts are arid and when this aridity set in, it is useful to consider where the world's major deserts are presently situated. The distribution of the world's major deserts ([Chapter 1](#), [Figure 1.1](#)) is closely linked to latitude and to distance from the sea. The Saharan and Arabian deserts, which extend eastwards across the deserts of Iran, Afghanistan and Pakistan into the Thar Desert of India, lie on or close to the Tropic of Cancer. The deserts of Australia, the Kalahari and the Atacama are traversed by the Tropic of Capricorn. On the other hand, the deserts of central Asia, including the Taklamakan and Gobi deserts of China and Mongolia, are situated in the interior of mid-latitude continental regions. A number of deserts are also located in the rain shadow of high mountain ranges, such as the Andes, the Rockies, the Himalayas and the Altai, Tian Shan and Kunlun ranges in central Asia. Why is this so?

Two sets of factors are responsible for this very particular distribution pattern. One involves the tectonic events that culminated in the global cooling and desiccation of the Cenozoic, a topic discussed in the next chapter ([Chapter 3](#)); the other is bound up with the present global atmospheric circulation system, itself a product of Cenozoic and earlier tectonic history. If we accept Lyell's dictum that in matters geological the present is the key to the past, it is logical to begin with the causes of present-day aridity before turning to the evidence for past aridity.

2.2 Causes of present-day aridity

There are four main reasons why deserts are lacking in rain (Mabbutt, 1977; Cooke et al., 1993; Thomas, 1997; Laity, 2008; Parsons and Abrahams, 2009; Nicholson, 2011). The two most important factors are location in a latitude dominated by dry subsiding air and location inland far from sources of moist maritime air. The remaining two factors are location in the rain shadow of high mountain ranges and location on a coast flanked by cold ocean currents or cold upwelling ocean water.

The first factor is a direct product of the global atmospheric circulation system, which is determined by solar radiation modulated by latitude, the distribution of land and sea, and the topography of the land. Solar radiation is greatest at the equator because the sun is most directly overhead there for much of the year. Away from the equator, progressively more incoming radiation is reflected or absorbed by the earth's atmosphere, because the sun's rays travel ever more obliquely through the atmosphere as a result of the curvature of the earth. Because of the tilt of the earth's axis, the sun is directly over each of the tropics once a year – at the summer solstice. If the axial tilt were greater, the sun would appear to travel further from the equator during summer, and the converse would apply if the tilt were less.

The tropical anticyclonic deserts, such as the Sahara, are a direct result of the atmospheric circulation cells (often termed *Hadley cells*) located between the equator and the tropics (Figure 2.1), so their location is determined by latitude rather than by the regional distribution of land and sea. For a full discussion, see Webster (2004). The two polar deserts also come under the influence of semi-permanent anticyclones and of cold, dry subsiding air. Because the distribution of high pressure cells (anticyclones) is closely related to latitude, the oceans in both polar and strictly tropical latitudes receive very little precipitation and are the arid marine counterparts of the continental deserts.

In equatorial latitudes, incoming short wave solar radiation heats both land and sea throughout the year. The air above both land and sea is constantly warmed by convection and outgoing long-wave terrestrial radiation. As the air becomes warmer, it moves upwards by convection, expands and eventually cools *adiabatically*, that is, without heat entering or leaving the air mass. The moist *adiabatic lapse rate* is about 5°C/km (3°C/1,000 feet), so the air mass will be 10°C cooler after an ascent of 2 km. Given that warm air can store more water vapour than an equivalent volume of cold air, the rising air soon reaches dew point, that is, it becomes saturated with respect to water vapour and excess water vapour condenses to form clouds. Cumulonimbus clouds can attain a thickness of 3–5 km, which means that during the rainy season equatorial regions are constantly cloudy, in contrast to the deserts, where there is little or no respite from the sun. Convictional uplift induces further cooling, leading to additional condensation of water droplets that eventually coalesce into larger drops and fall as

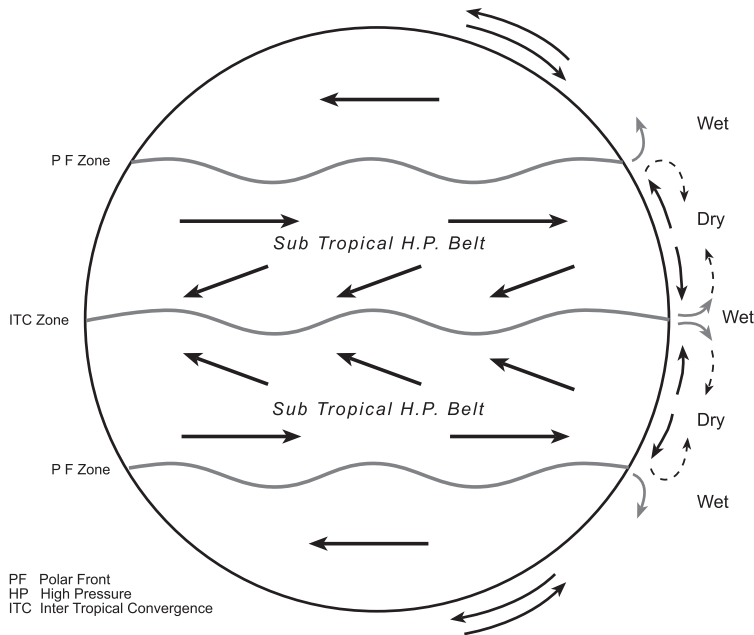


Figure 2.1. Schematic cross section of the global atmospheric circulation, showing location of the Hadley cells.

the heavy convectional downpours so characteristic of the wet and seasonally wet tropics.

Meanwhile, the air aloft becomes colder and denser and moves further away from the equator, until it finally begins to subside (Figure 2.1). The tropical latitudes centred on about 20–30° north and south are zones dominated by atmospheric subsidence. Depleted of much of its excess water vapour, the air over the two tropics is dry during the winter months but may become moist for a few months during the seasonal passage of the sun overhead and the attendant displacement of moist tropical air masses, known as the movement of the *Intertropical Convergence Zone*, or ITCZ (Figure 2.2). The further the distance from the equator, the shorter the tropical wet season becomes, until it ceases altogether and the deserts take over from the tropical savannas. Hot tropical deserts like the Sahara and Arabia are in latitudes where the air aloft is dry and subsiding and the atmospheric pressure is high for much of the year. The surface winds in deserts are therefore generally directed outwards, towards areas of lower atmospheric pressure, so there is minimal inflow of moisture from surface winds. As the air over the deserts subsides, it is compressed and becomes warmer, so that its capacity to absorb additional water vapour is increased. The result is that the *relative humidity* of desert air is usually very low and only reaches dew point (100 per cent relative humidity) when the night temperatures fall sufficiently for desert dew to precipitate

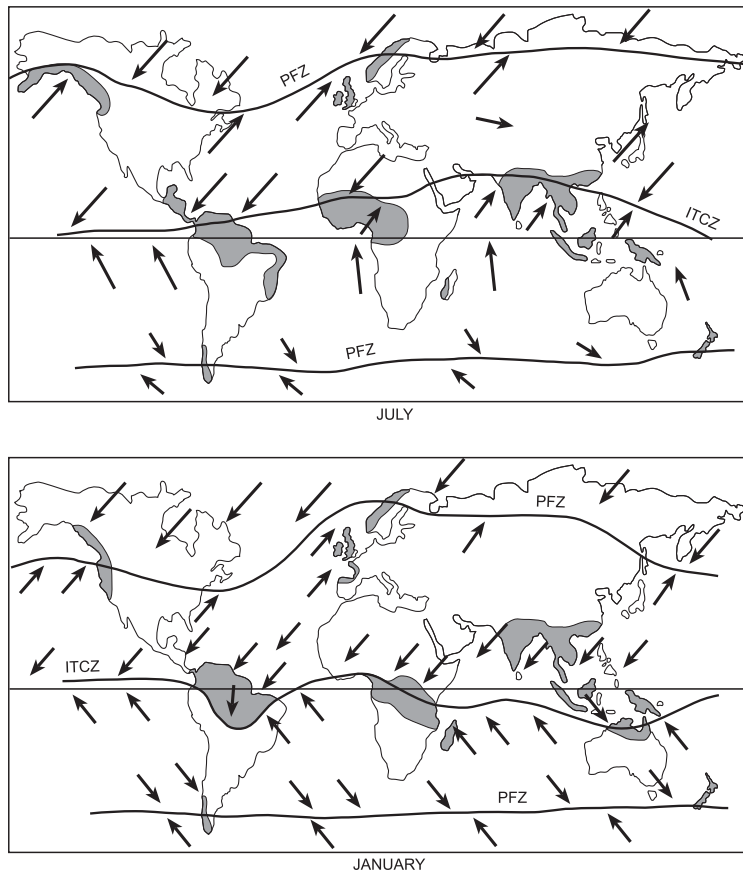


Figure 2.2. Seasonal migration of global wind systems and of the Intertropical Convergence Zone (ITCZ). (Adapted from *The Times Atlas of the World, Comprehensive Edition*, 1980.)

on chilled rock surfaces, especially the darker ones. This ephemeral dew allows some desert antelopes and other small creatures to survive despite the lack of surface water. Shortly before sunrise, small herds of gazelle may sometimes be seen licking the dew from the surface of small piles of desert rocks in North Africa and Arabia, a behavioural strategy that enables them to survive in otherwise waterless conditions.

The second major cause of aridity is a geographical location sufficiently far inland to be away from the influence of moist maritime air masses. Rainfall decreases rapidly away from the coast in all parts of the world except those close to the equator. Distance inland is sometimes described as *continentality* and naturally applies to all big deserts, including the great tropical deserts of Arabia, Australia and the Sahara. In the case of these hot tropical deserts, the effects of continentality accentuate those of latitude. Other examples of continental interior deserts are the great mid-latitude deserts of central Asia, Mongolia and western China. Bitterly cold in winter, with temperatures

falling as low as minus 40°C, they can experience summer temperatures close to 50°C, which is nearly as hot as the southern Libyan Desert in July and August.

Two other factors may either enhance the aridity resulting from latitude and continentality or may be the direct and dominant cause of reduced precipitation. These two factors are the proximity of cold oceanic water immediately offshore and the rain-shadow effect generated by high mountains. They may operate individually or together.

The presence of cold upwelling water or a cold ocean current close offshore is an effective cause of coastal aridity in tropical and even in equatorial latitudes such as the arid Horn of Africa, flanked by the cold Somali current. The cold Peru/Humboldt Current flows north parallel to the coast of the Atacama Desert in northern Chile and the coastal desert of Peru, with the cold California Current as its Northern Hemisphere counterpart, bringing aridity to Baja California. The cold Benguela Current flows north parallel to the Namib Desert in southern Africa, and the cold Azores Current accentuates the aridity of the western Sahara. The cold West Australian Current likewise flows north parallel to the arid west coast of Australia, but the situation here is more complex, with the warm Leeuwin Current flowing somewhat erratically from the Indonesian Warm Pool to the north to counteract the desiccating effect of the cold West Australian Current.

In fact, the western borders of all the great tropical or Trade Wind deserts in both hemispheres are washed by cool ocean currents associated with the oceanic circulation cells or gyres which flow clockwise in the northern hemisphere and anticlockwise in the southern hemisphere. If cool moist maritime air blows onshore, it often meets a land surface that is warmer than the adjacent ocean surface, at least in summer and during the day. The cool maritime air mass becomes warmer on contact with the warm surface of the land. The relative humidity of this air mass is therefore decreased, and its ability to absorb additional moisture from surface evaporation is increased. The air therefore has a desiccating effect on the land. This situation is only reversed if the land temperatures become significantly cooler than those of the adjacent ocean or if the sea surface temperatures become periodically warmer, as happens off the coast of Peru during El Niño years (see [Chapter 23](#)). Otherwise, the major sources of moisture in these often quite narrow coastal deserts are the coastal fogs that blow inland in winter when the land has cooled down relative to the sea surface temperatures. Coastal fogs are quite common in deserts where mountain ranges like the Andes or the Rockies, or uplands of more moderate elevation like the Red Sea Hills, run parallel to and close to the shore. For example, Erkowit in the Red Sea Hills of the eastern Sudan is a mist oasis and supports a spectacular flora of tall *Euphorbia candelabra* trees in the dry valleys between its rocky granite hills.

The fourth and final general cause of aridity is the rain-shadow effect, which is a global phenomenon linked to topography and is not restricted to deserts. Wherever ranges of hills or mountains are located close to the coast, forming a physical barrier

to onshore winds, the incoming moist maritime air will be forced upwards. As noted earlier, moist air is cooled adiabatically as it rises, attains vapour saturation and sheds its condensed water vapour as rain or snow. The air then passes over the coastal ranges and flows downhill, becoming warmer and drier. The country inland of the coastal ranges is described as being in the rain shadow of the ranges. The inland-facing slopes of high mountains are almost invariably drier than the coastal foothills, hence, for example, the great aridity of the Tibetan Plateau in contrast to the extreme wetness of the Assam foothills of the Himalayas. The wind-swept upland plains of the Bolivian Altiplano and of Patagonia lie in the rain shadow of the Andes. The rain-shadow deserts of New Mexico and Arizona are situated downwind of the Rockies. Other examples are the very gently undulating, semi-arid western plains of Queensland and New South Wales inland of the Eastern Highlands of Australia and the exceptionally hot, dry and rugged Afar Desert bounded by the Ethiopian Highlands to the south and west. Both the Afar Depression and the Dead Sea Rift are flanked by very high, mountainous escarpments and occupy low-lying fault-troughs or rifts that in places descend more than 150 m below sea level. If the region inland of the coastal uplands contains high mountains, these will be a focus for some additional *orographic*, or relief, rain, but if the area is lacking in relief, there will be no opportunity for any such precipitation. The gravel plains (*serir*) of the southern Libyan Desert and the sandstone plateaux (*hamada*) of the central Sahara, as well as the *gobi* plains of northern China and Mongolia and the *gibber* plains of central Australia, are all good examples of hyper-arid environments with very little surface relief. On these stony desert surfaces, rainfall and run-off are at a minimum and plants and animals are exceedingly rare, even by desert standards.

2.3 Evidence of formerly wetter climates in now arid areas

After considering why deserts are arid, it is pertinent to ask whether they have always been so. The answer is unequivocally no. Scattered across the Sahara Desert are the silicified trunks of tall trees that once grew in abundance in the forests that covered this region more than 100 million years ago. The prehistoric hunters who roamed the Sahara at intervals during the last million years made good use of this fossil wood to fashion the stone tools they used to hunt the great herds of savanna animals that also inhabited this now empty region. Later still, Neolithic pastoralists grazed their brindled herds of cattle, sheep and goats at numerous localities throughout the Sahara. They left behind them an enduring legacy of rock paintings and engravings on the smooth rock faces of the Tassili sandstone plateau in Algeria, the granite slopes of the Aïr Mountains in Niger and the sandstone plateaux and granite massifs of the Libyan and Egyptian deserts (Muzzolini, 1995; Coulson and Campbell, 2001).

Another indication of formerly wet climates in the drier parts of Africa, Asia and Australia is the ubiquitous presence of deeply weathered and chemically altered

bedrock. To achieve such a degree of intense leaching and new mineral formation requires considerable rainfall, a relatively dense vegetation cover and very low rates of physical denudation – conditions more reminiscent of the wet tropics than of the arid tropics.

Perhaps the most striking evidence of previously wetter conditions are the remains of the once integrated river systems that used to flow through every major modern desert, providing the sandy alluvium that was later fashioned by wind into the imposing sand seas and associated dunes popularly considered synonymous with deserts. Indeed, one of the most characteristic features of all deserts is their current lack of a perennial and integrated system of drainage (Cooke et al., 1993). Desert streams are ephemeral. They flow episodically, for variable distances, depending on the intensity and duration of sporadic rainstorms in their upper catchments. Even great rivers that flow through deserts, like the Nile, the Tigris and the Euphrates, originate in well-watered uplands far beyond those deserts. All rivers that flow through deserts constantly lose water by evaporation and by seepage to the local aquifers. Most desert rivers never reach the coast and instead flow into closed depressions, like the Tarim Basin in China or the Lake Eyre Basin in Australia. Such rivers are termed *endoreic*, in contrast to *exoreic* rivers like the Nile, which flow to the sea (de Martonne and Aufrère, 1928).

The fossil river valleys of the Sahara, the Gobi and western Australia have long provoked the curiosity of geologists (Chapter 10). Today they are broad, linear depressions filled with Cenozoic alluvium that is often cemented with iron, silica or calcium carbonate. Some of these former valleys now show relief inversion and form low erosional remnants or even extensive sheets of resistant ferricrete, silcrete or calcrete (see Chapter 15). In Mauritania, Namibia and western Australia, these valley-fill calcretes may also contain variable amounts of secondary uranium minerals precipitated out of slowly moving groundwater originating from the Precambrian host rocks that form the valley interfluves.

2.4 Evidence of previously greater aridity and desert expansion

Just as now arid areas retain evidence of once wetter climates, so is the converse equally true. Along the now vegetated and stable margins of all the great deserts, there is abundant evidence of former aridity in the shape of presently vegetated and stable desert dunes (Chapter 8), salt lake and evaporite deposits (Chapters 11 and 12), and vegetated mantles of desert dust (Chapter 9). Some care is needed in using such evidence to reconstruct past aridity, particularly in the case of desert dunes.

Desert dunes presently occupy about one-fifth of the Sahara and nearly two-fifths of the Australian arid zone. In North Africa, the 150 mm isohyet is a good indicator of the boundary between active and vegetated dunes, although precipitation is not the only factor responsible for dune mobility. Sand supply, wind velocity, surface roughness

and evaporation rates will also have an important influence on the movement of sand grains, as the presence of active coastal dunes in relatively wet areas with strong winds and abundant sand attests. The source-bordering dunes that form downwind of sandy channels in semi-arid areas provide another exception to the general rule that dunes cease to be mobile once the rainfall exceeds about 150 mm. As we shall see in more detail in [Chapter 8](#), there are three main prerequisites for the formation of source-bordering dunes. First is a regular, usually seasonal replenishment of river channel sands or of sandy beaches by long-shore drift in deep lakes. Second is a strong seasonal unidirectional wind, and third is a lack of riparian or lake-margin vegetation. The first prerequisite, a regular renewal of the sand supply from seasonally active rivers, precludes a fully arid climate.

Bagnold's classic observations on the Libyan Desert and his detailed experimental work (Bagnold, 1941) demonstrated that the volume of desert sand transported by wind increases exponentially with wind velocity above a certain threshold value, a finding confirmed by later workers (Pye and Tsoar, 1990; Cooke et al., 1993; Lancaster, 1995; Warren, 2013). Where sand supply and wind velocities are not limiting factors, dune mobility will increase as vegetation cover decreases. Rainfall and evapotranspiration are the primary controls over plant cover in arid areas. There is therefore a close relationship between the amount of rainfall and the average outer limit of active dunes in such deserts as the Thar or the Sahara. The belts of fixed and vegetated dunes along what are now the semi-arid margins of these two deserts have been mapped in detail from air photographs and satellite imagery. Assuming that the relationship between rainfall and dune mobility held good in the recent past, then the presence of these fixed dunes indicates that the effective range of the Sahara once extended 400 to 600 km further south and that of the Thar Desert some 350 km further south-east. [Chapter 8](#) enlarges on these general propositions.

2.5 Conclusion

Four main factors are responsible for aridity. The first factor is latitude. The hot tropical deserts are located in latitudes characterised by dry subsiding air. The reason for this is linked to the global atmospheric circulation system in which solar heating in equatorial latitudes causes air to rise and move towards the poles. The moist air aloft cools and sheds much of its water vapour as rain that falls over the equator and the adjacent seasonally wet tropics. As the air aloft continues its path towards the poles, it becomes cooler and denser and starts to subside between latitudes 25° and 30° north and south, creating zones of high pressure known as anticyclones. As it subsides, the air becomes warmer and its relative humidity decreases, so it has a desiccating effect on the land below. The second factor causing aridity is distance inland, an effect known as 'continentality'. Except in equatorial latitudes, the greater the distance from the coast, the lower the rainfall. The third factor is the presence

of cold ocean water immediately offshore, either because of cold upwelling ocean water or because of a cold ocean current flowing parallel to the coast. The fourth factor is location in the lee of a high mountain range running parallel to the coast, an effect termed the rain-shadow effect, because most of the moist air flowing in from the coast loses much of its moisture as it rises and cools on the windward side of the ranges.

At intervals in the past, changes in global atmospheric circulation have occurred at time scales ranging from decades to millennia, causing the deserts and their margins to become more or less arid. Evidence of past desert expansion is seen in the vast areas of now fixed and vegetated desert dunes that lie hundreds of kilometres beyond the margins of great deserts like the Sahara or in the now stable and soil-covered expanses of desert dust in central China. During episodes in the past when the deserts were less arid than they are today, integrated river systems flowed across them, lakes were abundant and fresh, and prehistoric human populations were able to graze their herds in areas now devoid of both water and pasture. In short, in some presently humid areas there is evidence of former aridity, just as some previously wetter areas are now arid.

3

Cenozoic evolution of deserts

A story that begins with tectonic uplift in the tropics and sub-tropics
thus ends with glaciation of the polar regions.

William F. Ruddiman

Tectonic Uplift and Climate Change (1997, p. ix)

3.1 Introduction

The Cenozoic era spans the last 65.5 million years (65.5 Ma) of geological time ([Table 3.1](#)) and follows the Mesozoic era that saw the proliferation and eventual extinction of the dinosaurs. It is in the Cenozoic that the mammals flourished and, in the last 7 million years, the African late Miocene and Pliocene hominids appeared that ultimately evolved into our Pleistocene human ancestors ([Chapter 17](#)).

The Quaternary Period, now internationally defined as beginning 2.58 million years ago (Gibbard et al., [2010](#)), comprises the Pleistocene Series/Epoch (2.58 Ma to 11.7 ka) and the Holocene Series/Epoch (11.7 ka to present) ([Table 3.2](#)). Plans are afoot to subdivide the Pleistocene into four Stages or Ages, namely the *Gelasian* (2.60 Ma to 1.80 Ma), the *Calabrian* (1.80 Ma to 780 ka), and, more tentatively, the *Ionian* (780 ka to 125 ka) and the *Tarentian* (125 ka to 11.7 ka). The Holocene ‘remains as a series/epoch distinct from the Pleistocene, in recognition of the fundamental impact of humans on an otherwise unremarkable interglacial’ (Gibbard et al., [2010](#), p. 101). However, a recent discussion paper suggests that the Holocene be formally subdivided into Early, Middle and Late, with the respective boundaries between Early and Middle at 8.2 ka and between Middle and Late at 4.2 ka (Walker et al., [2012](#)). There are proposals mooted to call the most recent slice of geological time the *Anthropocene*, in deference to our growing impact on the earth, but because human impact across the planet is highly time-transgressive, reaching agreement on when this proposed epoch might have begun will be far from easy, even assuming that the term serves a useful purpose.

Table 3.1. *Cenozoic time scale (From Geological Time Scale, Geological Society of America, 2009)*

Quaternary	0–2.6
Holocene	0–0.01
Pleistocene	0.01–2.6
Neogene	2.6–33.9
Pliocene	2.6–5.3
Miocene	5.3–23.0
Oligocene	23.0–33.9
Paleogene	33.9–65.5
Eocene	33.9–55.8
Palaeocene	55.8–65.5

In discussing the links between continental drift and mountain building, Arthur Holmes was moved to describe the earth as ‘an extremely old rotating electro-magnetic hydro-dynamic machine with a geochemical structure of great complexity’ (Holmes, 1965, p. 1247). He also speculated that there was a direct causal link between late Cenozoic mountain building and the inception of glaciation. Flint (1971) was more circumspect in his search for the causes of Quaternary glaciation and concluded that uplift alone was insufficient to trigger an ice age. It was not until a quarter century later that Ruddiman and Raymo (1988) and Ruddiman et al. (1997) were able to demonstrate in convincing detail the nature of the links between mountain building, weathering, erosion and climate change. The aims of this chapter are to give a concise overview of the Cenozoic record of desert evolution and to show the links between tectonic events, Cenozoic cooling and climatic desiccation.

3.2 Cenozoic tectonism, cooling and desiccation

The cooling and desiccation that characterise the second half of the Cenozoic and gave rise to the deserts we know today were a direct result of plate tectonic movements that

Table 3.2. *Quaternary time scale (ka BP)*

Holocene	0–12 ka
Upper Pleistocene	12–125 ka
<i>Tarentian</i>	
Middle Pleistocene	125–780 ka
<i>Ionian</i>	
Lower Pleistocene	780–2,580 ka
<i>Calabrian</i>	780–1,800 ka
<i>Gelasian</i>	1,800–2,600 ka

Table 3.3. *Global environmental changes of the past 130 million years*

TECTONICS, CLIMATE, VEGETATION: NORTHERN PERSPECTIVE
1. India-Asia contact: Himalayan uplift (~45 Ma)
2. East Africa: Uplift and rifting (~25 Ma)
3. Africa meets Europe: Mountain building (15 Ma)
4. Tethys Sea shrinks: Rainfall less and more seasonal
5. Forest gives way to savanna woodland and grassland in Africa and Asia
6. Messinian Salinity Crisis: Africa isolated from Eurasia (5.96–5.33 Ma)
7. Pliocene hominids appear in Africa (5 Ma)
8. Late Pliocene cooling and tropical desiccation (2.6 Ma)
9. First stone tools appear in Ethiopia: <i>Homo habilis</i> (2.5 Ma)
TECTONICS, CLIMATE, VEGETATION: SOUTHERN PERSPECTIVE
1. India separates from Australia-Antarctica (~128 Ma)
2. Australia first separates from Antarctica (~90 Ma)
3. Tasmania-Antarctic Passage opens (~34 Ma)
4. Drake Passage opens (~30 Ma)
5. Antarctic cooling and thermal isolation
6. Ice cap growth in Antarctica (34 Ma)
7. Cooling of the Southern Ocean
8. Australia moving north into dry subtropical latitudes
9. Forest replaced by woodland in Australia
10. Miocene ice expansion in East Antarctica
11. Late Miocene ice sheet in West Antarctica
12. Woodland gives way to savanna grassland in Australia, South America and South Africa
13. Drainage disruption, desiccation and faunal extinctions in Australia
TECTONICS, OCEAN CIRCULATION, GLACIATION
1. Laurasia break-up: North Polar cooling (15 Ma)
2. Antarctic cooling and ice build-up (34 Ma)
3. Cold Antarctic bottom water (34 Ma)
4. North Atlantic deep water (12 Ma)
5. Closure of Panama seaway: Arctic ice (3.5 Ma)
6. North American ice cap build-up (2.5 Ma)
7. Quaternary low-amplitude, high-frequency 40 ka glacial cycles (2.5 Ma)
8. Quaternary high-amplitude, low-frequency 100 ka glacial cycles (~0.9 Ma)
9. Glacio-eustatic sea level fluctuations
10. Alternating glacial desert expansion and interglacial desert contraction

began early in the Mesozoic (Table 3.3; Figures 3.1 and 3.2). Disintegration of the Pangaea supercontinent and of the super-ocean Tethys began in the Jurassic around 180 Ma ago (Kearey and Vine, 1996; Williams et al., 1998, pp. 11–21). The break-up of Pangaea was associated with higher levels of igneous activity along developing rift systems and the development of new continental margins. Early rifting led to the initial opening of the proto-Atlantic Ocean, with the emergence of the other oceans during the Mesozoic, culminating in the formation of the Southern Ocean early in the Cenozoic, from about 50 Ma onwards (Table 3.3). Rifting in East Antarctica at

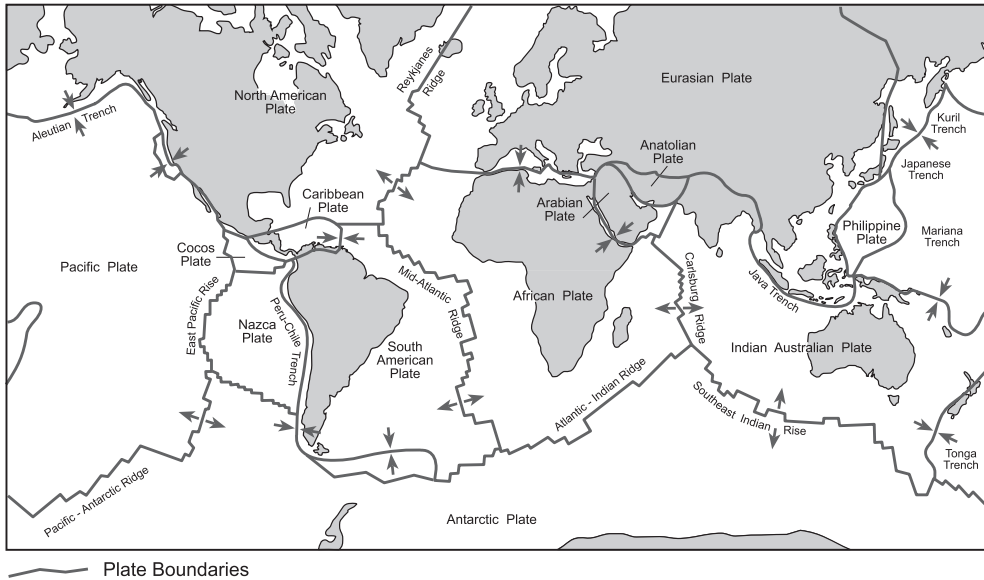


Figure 3.1. World tectonic plate boundaries. (After Williams et al., 1998, fig. 2.1.)

approximately 100 Ma led to the creation of a 2,500 km long rift system similar to that in East Africa today (Ferraccioli et al., 2011). Uplift along the rift margins created the Gamburtsev Mountains, which were deeply eroded by rivers and later by ice, and formed a nucleus for the build-up of the Antarctic ice sheet around 34 Ma and possibly for earlier ephemeral ice sheets.

Separation of Greater India from the western margin of Australia-Antarctica began some 128 Ma ago and was followed by its northward movement and eventual collision with and subduction beneath the Eurasian plate, leading to uplift of the Himalayas and of the Tibetan Plateau from about 45 Ma onwards (Table 3.4). This uplift in turn created a pronounced rain shadow to the north and led to accentuated aridity in the Taklamakan Desert of western China and the Gobi Desert of Mongolia. Late Cenozoic uplift linked to plate movements gave rise to the Andes in South America and the Rocky Mountains in North America, in turn creating rain shadows to leeward of both ranges. Northward movement of the African plate and its encounter with the Eurasian plate led to uplift of the Atlas Mountains, which formed a barrier denying access to the developing Sahara from moist northerly air masses.

Uplift of the Andes, the Rockies, the Sierra Nevada, Ethiopia-Arabia, the Atlas, the Alps and the Himalayas-Tibetan Plateau would have accelerated weathering and erosion, leading to an increase in the dissolved and suspended load carried to the oceans (Figure 3.3). (The timing of this uplift varied from region to region and is covered in Chapters 18 to 22, so we do not need to dwell on it further in this chapter). Roughly nine-tenths of the chemical and suspended loads in the Amazon today come

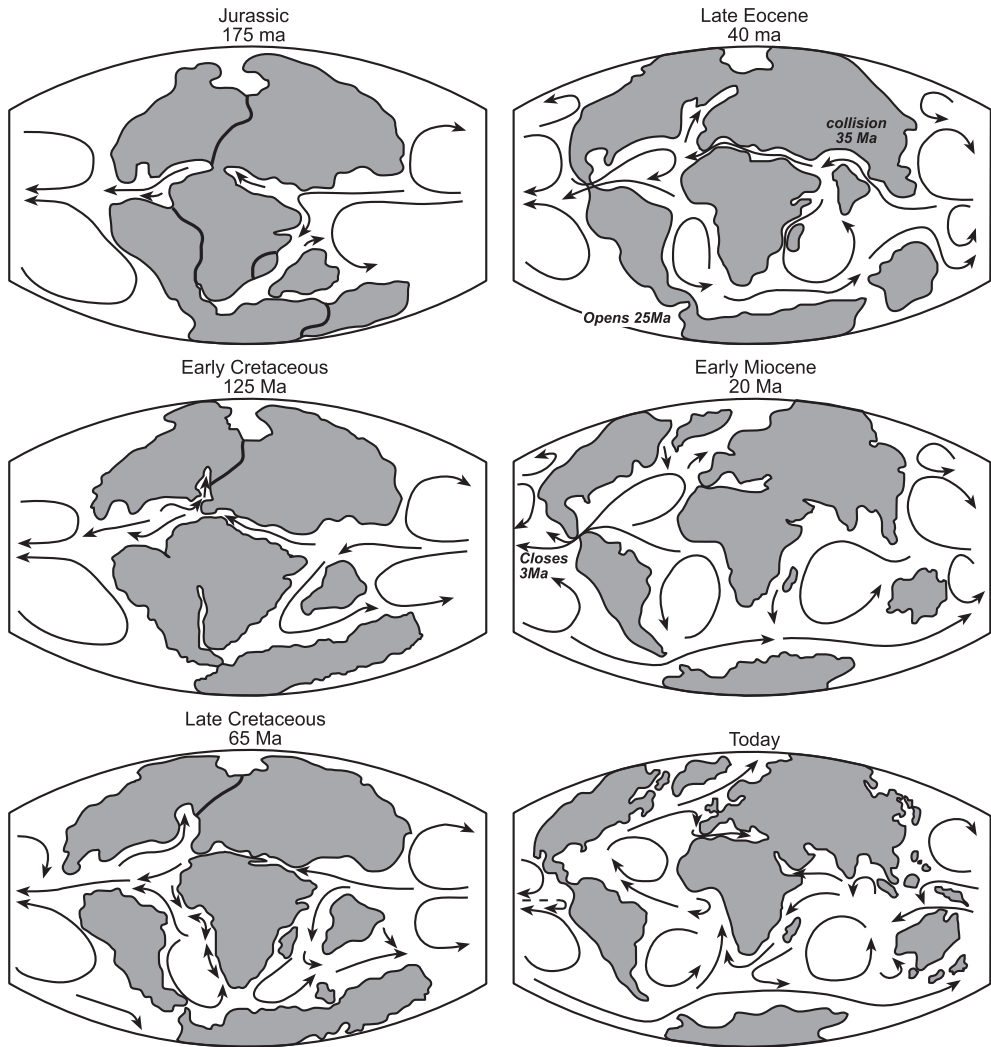


Figure 3.2. Cenozoic plate movements. (After Williams et al., 1998, fig. 2.2.)

from about one-tenth of the basin area, namely the mountainous headwaters, and the same is true of other major river systems originating in recently uplifted mountain catchments (Gibbs, 1967; Milliman and Meade, 1983; Milliman, 1997; Inam et al., 2007; Meade, 2007; Singh, 2007). Continued weathering would cause a slow but progressive removal of carbon dioxide from the atmosphere, promoting cooling of the lower atmosphere.

The separation of Australia from Antarctica and its northward movement at the fingernail growth rate of 5–6 cm/year brought Australia into subtropical latitudes characterised by dry subsiding air. The eastern margin of Australia moved across one or more stationary litho-thermal plumes, or hotspots, giving rise to volcanic activity

Table 3.4. *Late Cenozoic tectonic and climatic events*

50–45 Ma	Separation of Australia from Antarctica; inception of Southern Ocean
45 Ma	Northward movement of Australia into dry subtropical latitudes
	Collision of Greater India with Asia
	Progressive uplift of the Tibetan Plateau and Himalayas
38–30 Ma	Dropstones in Norwegian-Greenland Sea
	Ice present in Greenland?
34–33 Ma	Opening of Drake Passage between Antarctica and South America
	Creation of circum-Antarctic Current
	Major ice accumulation in Antarctica
	Major global cooling
6–5 Ma	Severe desiccation in central Asia
	Miocene salinity crisis
	Mediterranean salt desert
	Incision of Nile canyon
	Genetic isolation of Africa from Eurasia
4–3 Ma	Emergence of bipedal hominids
	Closure of the Indonesian Seaway
	Diversion of cool ocean water towards East Africa
	Desiccation in East Africa
	Closure of the Panama Isthmus
2.7–2.5 Ma	Rapid accumulation of ice over North America
	Drying out of East Africa and the Sahara
	First appearance of stone tool-making in East Africa
	Enhanced aridity in central Australia
2.4–0.9 Ma	High-frequency, low-amplitude 41 ka glacial-interglacial cycles
0.9–0 Ma	Low-frequency, high-amplitude 100 ka glacial-interglacial cycles

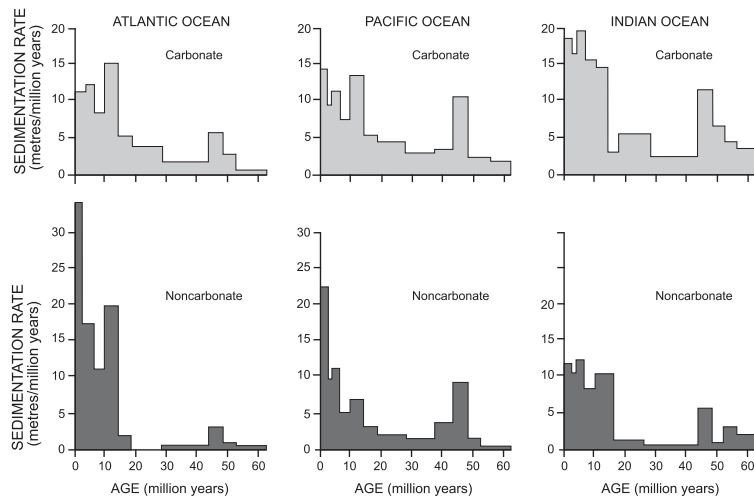


Figure 3.3. Cenozoic sedimentation in the Atlantic, Pacific and Indian Oceans. (After Williams, 2012, constructed from data in Davies et al., 1977.)

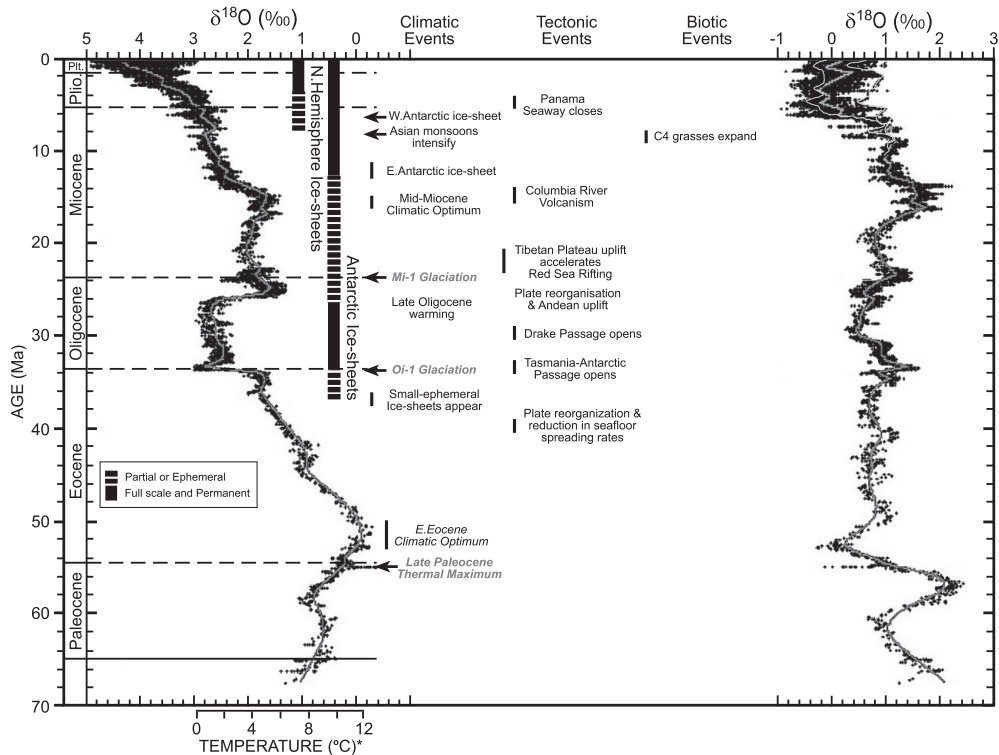


Figure 3.4. Major tectonic and climatic events of the Cenozoic. (Modified from Zachos et al., 2001, fig. 2.)

and further sporadic uplift of the Eastern Highlands, which accentuated the rain shadow west of the divide, increasing aridity over central Australia (Table 3.3).

Accumulation of continental ice in Antarctica may seem somewhat remote from desiccation in Africa and Australia but was in fact of critical importance. Mountain glaciers were present on Antarctica late in the Eocene, and a large permanent ice cap was established at the Eocene-Oligocene transition 34 Ma ago (Zachos et al., 2001) (Figure 3.4). Continental ice was slower to form in the Northern Hemisphere (Table 3.3) but was present in high northern latitudes by 3 Ma, or possibly by 5 Ma or even earlier, with a rapid increase in the rate of ice accumulation around 2.5 Ma (Shackleton and Opdyke, 1977; Shackleton et al., 1984; Clark et al., 2006).

As temperatures declined over the poles and sea surface temperatures at high latitudes grew colder, the temperature and pressure gradients between the equator and the poles increased. There was a corresponding increase in Trade Wind velocities and hence in the ability of these winds to mobilise and transport the alluvial sands of the Saharan, Kalahari and Australian depocentres and to fashion them into desert dunes. Higher wind velocities were also a feature of glacial maxima during the Pleistocene and were responsible for transporting Saharan desert dust far across the Atlantic (Parkin and Shackleton, 1973; Parkin, 1974; Williams, 1975; Sarnthein, 1978;

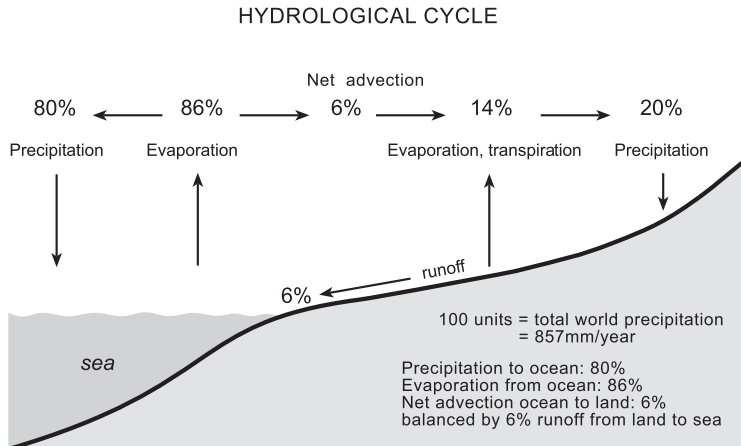


Figure 3.5. The global hydrologic cycle (schematic).

Sarnthein et al., 1981; Maher et al., 2010). Before and during the Last Glacial Maximum, centred on 21 ± 2 ka (Mix et al., 2001), Australian and Patagonian desert dust was also blown as far as Antarctica (Petit et al., 1981; Harrison et al., 2001; Revel-Rolland et al., 2006; Maher et al., 2010). In addition, stronger Trade Winds linked to a more intense Hadley circulation would have accentuated upwelling of cold ocean water on the upwind sector of the ocean gyres, strengthening the cold Benguela Current off Namibia and the cold Humboldt Current off Chile and Peru. In the Northern Hemisphere, the corresponding cold currents are the Azores Current off the western Sahara and the California Current off Central and North America, both of which contribute to enhanced aridity onshore.

Late Cenozoic cooling of the ocean surface was an additional factor responsible for reducing intertropical precipitation. Galloway (1965a) estimated that two-thirds of present-day global precipitation falls between latitudes 40°N and 40°S and depends significantly on evaporation from the warm tropical seas. In this context, Figure 3.5 shows the importance of the oceans in contributing water to the land. The ocean surface cooling, which was linked to global cooling associated with high-latitude continental ice build-up and enhanced cold bottom-water circulation, would help reduce evaporation from the tropical seas, thereby further reducing rainfall across the tropical deserts. Table 3.4 provides a global summary of some key Cenozoic tectonic and climatic events, together with their impact on oceanic circulation.

3.3 Cenozoic desiccation was not synchronous

The onset of late Cenozoic aridity and the resultant slow emergence of the deserts were the results of global tectonic events that led to changes in global atmospheric circulation linked to changes in the global distribution of land and sea. For example, the origin of the Sahara as a desert was associated with several independent tectonic

events. Slow northward movement of the African plate during the last 100 million years (late Mesozoic and Cenozoic) resulted in the migration of North Africa from wet equatorial into dry tropical latitudes. A slight clockwise rotation of Africa began about 15 million years ago and continued through the Miocene and Pliocene, bringing Africa into contact with Europe. This displacement was accompanied by crustal deformation and rapid uplift in the Atlas region and by volcanic eruptions and gentle updoming in Jebel Marra (3,042 m), Tibesti (3,415 m), the Hoggar (2,918 m) and the Aïr Mountains. Owing to their altitude, the high mountains of the central and southern Sahara have always been wetter than the surrounding desert plains and so may have served as refugia for plants, animals and humans throughout the Quaternary.

Two additional factors were responsible for the late Cenozoic desiccation of the Sahara-Arabia and Asian deserts. One was the gradual expansion of continental ice in high latitudes, which was associated with the cooling of the Southern Ocean. The final separation of Australia from Antarctica some 45 Ma ago culminated in the establishment of a large ice cap on Antarctica by 34 Ma ago. The closure of the Panama Isthmus and diversion of warm water into the North Atlantic, in conjunction with high northern latitude cooling, provided the impetus for a sudden increase in the volume of Northern Hemisphere ice caps around 2.5 Ma ago.

One effect of the progressive build-up of high latitude ice sheets, noted in the previous section, was to steepen the temperature and pressure gradients between the equator and the poles, resulting in increased Trade Wind velocities. Faster Trade Winds were better able to mobilise the alluvial sands of an increasingly dry Sahara and to fashion them into desert dunes. For example, the first appearance of wind-blown quartz sands in the Chad Basin occurs towards the end of the Cenozoic, when they were interstratified among late Pliocene to early Pleistocene fluvial and lacustrine sediments (Servant, 1973; Servant and Servant-Vildary, 1980; Sepulchre et al., 2006). The associated lacustrine diatom flora indicates temperatures cooler than those now prevalent in this region. The combined evidence suggests that the late Pliocene was both cooler and drier along the tropical borders of the Sahara. The diatom and pollen evidence from a large late Pliocene lake at Gadeb in the south-eastern uplands of Ethiopia (Gasse, 1980; Bonnefille, 1983) is also consistent with the inference that intertropical cooling and desiccation may have been closely bound up with the expansion of Northern Hemisphere ice caps around 2.5 Ma ago.

A further factor contributing to the drying out of the Sahara and Arabian deserts was the late Cenozoic uplift of the Tibetan plateau and the ensuing creation of the easterly jet stream that brought dry subsiding air to the incipient deserts of Pakistan, Iran, Arabia, Somalia, Ethiopia and the Sahara. Isotopic analysis of fossil soils and fossil herbivore teeth collected from the Potwar Plateau of Pakistan indicates a major change in flora and fauna between 7.3 and 7 Ma ago. Until about 7.3 Ma ago, forest and woodland dominated the landscape. After 7 Ma, there was a rapid expansion of tropical grassland at the expense of the forest. This change in vegetation may indicate the inception (or strengthening) of the Indian summer monsoon 7 Ma ago.

(Quade et al., 1989). Other factors have probably contributed to intertropical cooling and desiccation during the past 30 Ma. One was the progressive shrinkage of the Paratethys Sea. This warm, shallow sea once stretched across Eurasia but shrank gradually during the Oligocene and Miocene. As the once extensive sea shrank, the rainfall that was previously well-distributed throughout the year became progressively more seasonal. A further agent of late Cenozoic cooling was the decrease in atmospheric carbon dioxide associated with increased erosion, weathering and associated consumption of carbon dioxide caused by the late Cenozoic uplift of the Himalayas, the Rockies, the Andes, the Ethiopian uplands and perhaps also the Transantarctic Mountains. The global increase in plants using C_4 photosynthesis and the reduction in C_3 plants between about 8 and 6 Ma ago (Quade et al., 1989) is certainly consistent with a decrease in the concentration of atmospheric carbon dioxide. The threshold for C_3 photosynthesis is higher at warmer latitudes, and so it is not surprising that the initial change from C_3 to C_4 plants occurred in the lowland tropics first. Climatic cooling was probably also triggered by the eruption of the voluminous Ethiopian flood basalts over a period of no more than 1 million years around 30 Ma ago (Pik et al., 2003; Pik et al., 2008).

Changes in the Cenozoic flora and fauna of the Sahara show a similar trend to that inferred for the Himalayan foothills of Pakistan. During the Palaeocene and Eocene, much of the southern Sahara was covered in equatorial rainforest, and there was widespread deep weathering at this time. During the Oligocene and Miocene, much of what is now the Sahara was covered in woodland and savanna woodland, but by Pliocene times many elements of the present Saharan flora were already present (Maley, 1980; Maley, 1981; Maley, 1996). Pollen preserved in scattered localities in northern Africa shows that the replacement of tropical woodland by plants adapted to aridity was already underway during the late Miocene and early Pliocene (Maley, 1980; Maley, 1981; Maley, 1996), a conclusion consistent with the pollen evidence preserved in deep-sea cores off the north-west coast of Africa (Leroy and Dupont, 1994; Leroy and Dupont, 1997).

From about late Pliocene times onwards, the great tropical inland lakes of the Sahara, Ethiopia and Arabia began to dry out. The formerly abundant tropical flora and fauna of the well-watered Saharan uplands became progressively impoverished as entire taxa became extinct, and a once integrated and efficient network of major rivers became increasingly obliterated by wind-blown sands. In the Chad Basin there is good evidence of wind-blown desert dune sands deposited between alluvial and lacustrine sediments that were laid down more than 2 million years ago (Servant, 1973). Further north, in the Tibesti and Hoggar mountains of the central Sahara, the evidence from fossil pollen grains shows that some of the plants growing in this region were already adapted to aridity at about the time that the desert sands made their first appearance in the Chad Basin (Maley, 1980; Maley, 1981).

In central China, the first appearance of wind-blown desert dust was initially dated to around 2.4 Ma ago (Heller and Liu, 1982), but further to the north-west in central

Asia it is far older, with strong evidence of aridity in that region by around 24 Ma (Dupont-Nivet et al., 2007; Dupont-Nivet et al., 2008; Sun et al., 2010). In central Australia, the inception of aridity seems to be far younger, with major inland lakes persisting until at least 1 million years ago before giving way to desert playa lakes and dunes fashioned from wind-blown quartz and gypsum sand-sized particles (Chen, 1989; Chen and Barton, 1991; Fujioka and Chappell, 2010). Further afield, in the present Congo/Zaire Basin of central Africa, there are much older deposits of red desert sands that predate the oldest sands of the present Kalahari Desert. We therefore need to allow for local differences in the timing of early desiccation that are linked to regional climatic and tectonic factors and should not expect a similar sequence of Cenozoic events in the deserts of China, India, Australia, southern Africa and the Americas.

3.4 Quaternary climatic fluctuations in deserts

Before we conclude this brief survey of Cenozoic cooling and desiccation, it is important to draw attention to a number of unique climatic changes that took place at the onset of the Pleistocene. The start of the Pleistocene epoch (2.6 Ma to 12 ka) was heralded by a number of important regional environmental changes, including a major expansion of Northern Hemisphere ice, renewed widespread accumulation of wind-blown dust in the Loess Plateau of central China and the onset of a cooler, drier climate in the Ethiopian Highlands of East Africa. It is probably not a coincidence that the first evidence of stone tool-making by ancestral humans in the Afar Desert of Ethiopia is dated to around 2.5 Ma – a time of abrupt and rapid environmental changes in that region (see [Chapter 17](#)). Closure of the Panama seaway at the end of the Pliocene would have diverted warm ocean water northwards into the North Atlantic, thereby providing a sufficient source of moist air from the ocean to feed the growing ice caps in North America, Greenland and north-west Europe.

Subdividing the Pleistocene has often engendered more heat than light, but a good case exists for using the Brunhes-Matuyama paleomagnetic boundary (dated to 0.78 Ma) to delineate the boundary between Lower and Middle Pleistocene (Pillans, 2003). This proposal also accords with a similar and much earlier recommendation in the volume edited by Butzer and Isaac (1975) that arose from a Burg Wartenstein symposium dealing with cultural change in the Middle Pleistocene. The Upper Pleistocene extends, somewhat arbitrarily, from the peak of the last interglacial to the start of the Holocene (i.e., from 125 ka to 12 ka). The Upper Pleistocene thus consists of a single interglacial-glacial cycle, culminating in the Last Glacial Maximum, dating to 21 ± 2 ka ago (Mix et al., 2001), when global ice volume was last at its maximum and sea levels were correspondingly low (–120 m: Yokoyama et al., 2000; Lambeck and Chappell, 2001). The most recent subdivision of the Quaternary is into four Stages or Ages and simply adds an earlier phase for what was once the late Pliocene. These

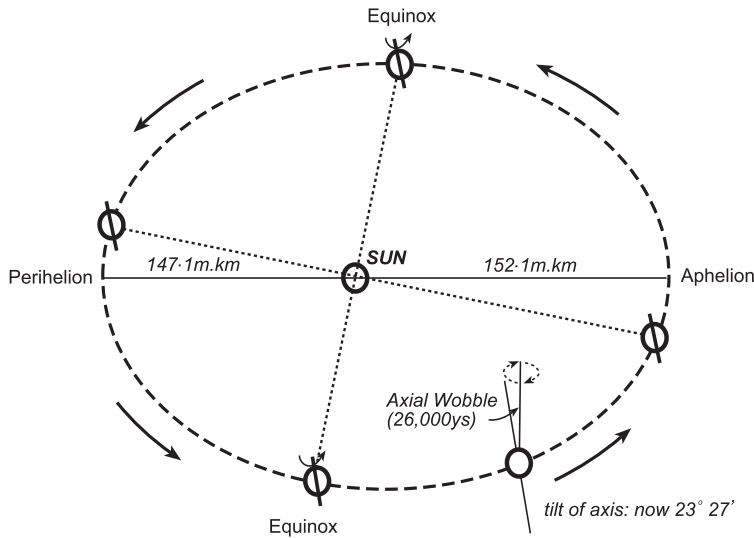


Figure 3.6. Orbital fluctuations.

ages are: 2.60 Ma to 1.80 Ma, 1.80 Ma to 780 ka, 780 ka to 125 ka and 125 ka to 11.7 ka (Gibbard et al., 2010).

3.4.1 Orbital periodicities during the Pleistocene

The history of the Pleistocene is one of alternating expansion and contraction of ice caps and of associated expansion and contraction of deserts, with concomitant changes in the rivers, lakes, glaciers and dunes in and around the deserts. The first two-thirds of the Pleistocene was a time of low-amplitude, high-frequency climatic fluctuations, with each glacial-interglacial cycle lasting about 40,000 years. The last third of the Pleistocene, starting about 0.7 Ma ago, was a period of high-amplitude, low-frequency climatic changes, with each glacial-interglacial cycle lasting about 100,000 years.

The amount of solar radiation received from the sun in any given latitude depends on three astronomically controlled variables (Milankovitch, 1920; Milankovitch, 1930; Milankovitch, 1941; Berger, 1981). One is the distance of the earth from the sun. The earth follows an elliptical path around the sun each year, with the sun not quite at the centre of the ellipse (Figure 3.6). At present, when the earth is closest to the sun (termed the perihelion), the distance is 147.1 million km. When furthest from the sun (termed the aphelion), the corresponding distance is 152.1 million km. The shape of the ellipse varies over time, being sometimes more elliptical and sometimes less so. This cyclical change in what is termed the orbital eccentricity has a duration of 96,600 years and is responsible for a 3.5 per cent variation in solar radiation received in the outer atmosphere.

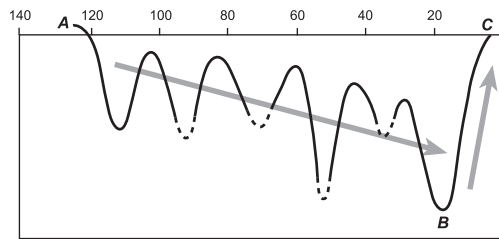


Figure 3.7. The last glacial-interglacial cycle (schematic). (A) is the last interglacial, (B) the Last Glacial Maximum and (C) the present interglacial. The sea level fluctuations associated with the expansion and contraction of the ice caps are simplified. Note the slow, saw-toothed build-up to full glacial conditions and the short duration of the last interglacial.

The inclination of the earth's axis (at present $23^{\circ} 27'$) also varies over time, from a maximum tilt of $24^{\circ} 36'$ to a minimum of $21^{\circ} 59'$. When the tilt is at a maximum, the summers in the higher latitudes tend to be hotter and the winters tend to be colder. During times of minimum tilt, summers are less hot and winters are milder (Figure 3.7). Mild cloudy summers offer a better chance for snow to persist in high northern latitudes than do hot dry summers. This obliquity cycle of 41,000 years is therefore the main control over seasonality (Williams et al., 1998).

The third cycle (often referred to as the precession of the equinoxes or, more simply, the precessional cycle) reflects the changing season of the year when the earth is nearest to the sun and is governed by the direction in which the spin axis of the earth points in space. The precessional cycle is quite variable in duration, with a mean period calculated by Milutin Milankovitch (1941) for the last 1 million years of 21,000 years, although he noted that it had ranged from 16.3 ka to 25.8 ka in that time. James Croll (1875) was aware of the role of the precessional cycle in altering the seasonal distribution of heat received at the earth's surface, but his ice age model, although prescient, failed to explain synchronous glacial and interglacial cycles in both hemispheres, and it postulated too short a duration for each cycle.

The dominant cycles evident in late Pliocene deep-sea cores up until 2.6 Ma ago were the 23 ka and 19 ka precessional cycles, after which the 41 ka obliquity cycles dominated until about 0.7 Ma ago, when the 100 ka orbital eccentricity cycles became dominant (Williams et al., 1998; Lisiecki and Raymo, 2005; Clark et al., 2006; Lisiecki and Raymo, 2007). The interval known as the 'Middle Pleistocene Transition' began about 1,250 ka ago and had ended by 700 ka ago, when the 100 ka cycles had become dominant (Clark et al., 2006; Lüthi et al., 2008).

3.4.2 Long Pleistocene records of climatic change

Very few terrestrial records span the entire duration of the Pleistocene. The long pollen records from the Bolivian Altiplano are one such archive. Another outstanding archive

is the Chinese loess record, which shows an alternation of loess accumulation in central China during cold, dry and windy climatic interludes and soil development under a re-established cover of moderately dense vegetation during the warmer, wetter intervals when the summer monsoon had become stronger once more. The dry intervals were coeval with glacial or stadial climatic phases and the wet intervals with interglacial or interstadial phases.

The ice core records from Antarctica and Greenland provide a second, highly informative, set of terrestrial archives. The Antarctic records from Vostok, near the centre of the ice cap, and EPICA Dome C now span the past 0.8 million years. They reveal that temperature over the ice cap fluctuated in parallel with the concentrations of atmospheric carbon dioxide ($p\text{CO}_2$) and methane. During glacial maxima and times of minimum local temperature, the $p\text{CO}_2$ levels were around 180–200 parts per million by volume (ppmv), rising to 280–300 ppmv during interglacials (Jouzel et al., 1997; Lüthi et al., 2008). The corresponding methane values were around 400 parts per billion by volume (ppbv) during glacials and 800 ppbv during interglacials (Petit et al., 1999). The dust concentration revealed in the ice cores also shows significant fluctuations, with peak concentrations of dust blown from Patagonia and arid Australia coinciding with glacial maxima.

Finally, the Mediterranean Sea contains a long and detailed record of past climates. During times of high fluvial discharge into the sea from the Nile and from now inactive Saharan rivers, dark, organic-rich sediments, or sapropels, were laid down on the floor of the sea (Rossignol-Strick et al., 1982; Larrasoana et al., 2003; Ducassou et al., 2008; Ducassou et al., 2009). The sapropel units are believed to reflect accumulation in anoxic bottom waters during times of enhanced freshwater flow into the Mediterranean (Scrivner et al., 2004; Tzedakis, 2009). The sapropels alternate with calcareous muds that contain a high proportion of Saharan wind-blown dust. The alternating sequence of sapropels and calcareous muds rich in eolian dust thus reflect times of stronger and weaker summer monsoons over northern Africa (Freydier et al., 2001; Ducassou et al., 2008; Ducassou et al., 2009; Tzedakis, 2009), as well as times of enhanced or decreased winter rainfall along the north coast of Africa (Rossignol-Strick, 1985).

3.4.3 Millennial-scale Pleistocene and Holocene climatic fluctuations

The Greenland ice core oxygen isotope records that span the last glacial-interglacial cycle reveal a pattern of frequent and rapid temperature changes. Dansgaard et al. (1993) found evidence of twenty-four warm *interstadial* events ranging from approximately 2,000 to approximately 500 years in duration. Each showed a rapid temperature increase of up to 7°C within just a few decades, followed by a slower cooling down to 12–13°C below modern levels (cold *stadial* events). Each of the packets comprising a warm interstadial and a cold stadial is known as a ‘Dansgaard-Oeschger’ (D-O) event, lasting 1,000–3,000 years (Williams et al., 1998). Heinrich (1988) noted periodic influxes of ice-rafted debris into the North Atlantic during the late Quaternary,

and these are now termed 'Heinrich events'. Bond et al. (1997; 2001) found that Holocene episodes of ice-rafted debris (IRD) into the North Atlantic ('Bond cycles') occurred at intervals of approximately 1,500 years, possibly linked to changes in solar activity.

3.5 Conclusion

Several immediate conclusions may be drawn from this brief survey. First, tectonic factors involving lateral migration of the continents, opening up of the oceans and Cenozoic uplift of major mountain ranges like the Himalayas, Rockies and the Andes were the primary causes of Cenozoic cooling and desiccation. The timing of these tectonic events varied from region to region, which means that the inception of aridity and the birth of the major deserts were not everywhere synchronous, suggesting that local influences need to be considered in any account of climate change in deserts. During the Quaternary period, which began 2.6 Ma ago, astronomically controlled changes in the amount of solar radiation received by the earth determined the duration and severity of the successive glacial-interglacial cycles. Over the past 0.7 Ma, the 100 ka cycle has been dominant, although the 20 ka precessional cycles have also had a major influence on climate. During times of maximum high latitude glaciation, sea surface temperatures were lower, intertropical rainfall diminished, rainforests shrank and the great tropical deserts in Africa, Arabia, India and Australia expanded. The reverse was true during interglacial and interstadial episodes, with re-integration of former drainage systems, desert lakes becoming abundant, and formerly active desert dunes and loess plains along the desert margins becoming vegetated and stable. It was during these occasions that prehistoric human groups were able to penetrate the deserts and travel across the presently forbidding deserts of the Sahara, southern Negev and Arabia. Finally, the onset of aridity in Africa, Asia, Australia and the Americas predates any human presence in those regions, and so humans cannot be considered the cause of these deserts, as some have erroneously claimed.

4

Adaptations to life in deserts

The nomads will not burn the good pasture bushes . . . even in their enemies' country. . . . I have sometimes unwittingly offended them, until I knew the plants, plucking up and giving to the flames some which grew in the soil nigh my hand.

Charles M. Doughty (1843–1926)

Passages from Arabia Deserta (1931)

4.1 Introduction

Over very long intervals of time, the plants and animals living in deserts and their margins have become well-adapted through their morphology, physiology and behaviour to using scarce water efficiently (Evenari et al., 1971; Stafford Smith and Morton, 1990; Morton et al., 2011). Evenari et al. (1971) noted that the more extreme the desert habitat, the more specific were the requirements for survival. The sparse human populations in deserts have also evolved long-term behavioural adaptations to the harsh extremes of desert climates, particularly through a nomadic lifestyle designed to make optimum use of sporadic rains and ephemeral grazing. The sedentary communities who live in deserts and depend on plant cultivation have learned to obtain and use water with great ingenuity, as in the case of the underground water-conducting tunnels of the piedmont deserts of the Near East, central Asia, the Sahara and Mexico. In those favoured desert localities where permanent springs exist, or where groundwater can be tapped by wells and by deep-rooted, relatively salt-tolerant trees such as date palms, larger human settlements become possible, but many oases are today facing problems of falling water-tables and increasing salt accumulation in soils used for growing crops.

4.2 Water in deserts

Without water life is not possible, and in deserts it is the availability of water that determines where plants, animals and humans can live (Figure 4.1). For plants, especially



Figure 4.1. Afar women filling goatskin waterbags, Afar Desert, Ethiopia.

the higher plants, water is *the* limiting factor, because in contrast to animals that can move to more favourable environments should the need arise, their mobility is restricted by their root systems. Successful adaptation to desert living thus requires an ability to make optimum use of the sporadic distribution of water in time and space (Evenari et al., 1971; Stafford Smith and Morton, 1990; Morton et al., 2011). As a general rule, the lower the total annual rainfall, the more variable it is from year to year, so that areas that normally receive little or no rain may receive several hundred millimetres of rain in a single, highly localised downpour, sometimes leading to severe flood damage and loss of life. Dorothea Mackellar (1885–1968) caught this distinction nicely in her poem *My Country* in which arid inland Australia is ‘*a land of sweeping plains, of ragged mountain ranges, of droughts and flooding rains*’. The contrast here between plain and mountain is important, because the distribution of surface and subsurface water is quite different in both zones. Indeed, rock type and relief exert a dominant control on water availability in deserts, at all scales from regional to local.

Nearly a century ago, the French geographers Emmanuel de Martonne and L. Aufrère (1928) classified drainage systems into three broad categories: *exoreic*, *endoreic* and *areic*. Exoreic river systems flow from their upland headwaters to the sea, with the Nile River being a well-known example. Endoreic river systems occupy internal drainage basins and fail to reach the sea, sometimes terminating in inland

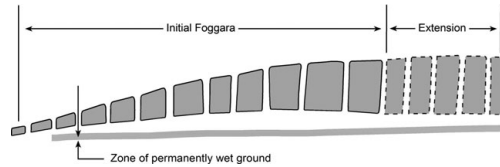


Figure 4.2. Cross section through a *foggara*, central Sahara. (After El Hadj, 1982.)

lakes, like Lake Chad or the Aral Sea. The Syr Darya and Amur Darya in central Asia are good examples of such drainage systems, with prolonged water abstraction from the Syr Darya for cotton irrigation being the main cause of the shrinking of the Aral Sea. Areic drainage systems are usually ephemeral and lack an integrated drainage network, although they may have possessed such a network under a previously less arid climate. The linear salt lakes of western Australia are the remnants of a once integrated and extensive drainage system that was active when Antarctica and Australia formed one large and well-watered continent more than 45 million years ago, but they now form part of an areic drainage system.

There are strong physiographic controls on the surface and shallow sub-surface distribution of water in deserts. Surface water is generally absent from dunes, gravel-covered plains and plateaux. A plethora of local names has been used to describe these landforms, with the gravel plains known as *reg* or *serir* in the Sahara, *gobi* in Mongolia and *gibber* plains in Australia. Likewise, the rocky plateaux are known as *hamada* in the Sahara, *mesas* in the American deserts and *stony tablelands* in Australia. Springs and ephemeral, seasonal or even perennial streams tend to occur within mountain valleys or along mountain fronts, while small ponds may form in the hollows between dunes after local heavy rain. In the Badain Jaran desert of Inner Mongolia, where the dunes attain relative heights close to 500 metres (Yang, 1991; Yang et al., 2011a), making them the highest dunes in the world, the depressions between the dunes are occupied by permanent lakes, although many of these lakes have been shrinking during the last few thousand years, most probably as a result of a progressively drier climate (Yang and Williams, 2003).

Alluvial fans at the foot of mountains have long been a major source of shallow groundwater for plants, animals and humans and some ingenious methods of water extraction, such as the *foggara*, *karez* or *qanats* discussed in section 4.5, are still in use today among peasant farmers in the semi-arid world (Figure 4.2). Shallow sub-surface water also occurs in the sand and gravel beds of ephemeral or seasonal stream channels and can be accessed from wells dug one or more metres below the surface. These temporary wells remain a major source of water for nomads and small villages throughout the semi-arid world today. Many permanent wells have been sunk away from the river channels to tap aquifers at greater depth. When dug through clay or sandy clay, they are usually lined with bricks or rocks. Several remarkable wells

more than 100 metres deep were dug by hand nearly two centuries ago through hard Nubian Sandstone to provide reliable water for humans and their herds during the sixteen-day journey by camel from Kufra Oasis in Libya to Ounianga in Chad. In fact, the Nubian Sandstone is the largest groundwater aquifer in the world (Alker, 2008, extending across 2 million km² in the Sahara and Arabia, with an estimated storage of 150,000 km³ of groundwater. Unfortunately, the water in this aquifer is being used today at a rate far in excess of any replenishment from precipitation on distant mountains. The last episodes of groundwater recharge of this aquifer were in the late Pleistocene and early Holocene, so current use amounts to mining a non-renewable resource.

A number of other major desert aquifers, such as the Ogallala Aquifer in the United States (Reilly and Franke, 1999; Bartolino and Cunningham, 2003) and the Great Artesian Basin in central Australia, are also being used to excess, which will require new adaptive strategies in the future. The Great Artesian Basin covers an area of 1.7 million km² (660,000 sq miles) and is the largest and deepest artesian basin in the world. (Artesian water is confined groundwater flowing under pressure from a recharge zone of higher elevation to an outlet in a series of springs situated at lower elevations). The Ogallala Aquifer, also known as the High Plains Aquifer, is a vast shallow aquifer some 450,000 km² (174,000 sq miles) in area that underlies eight states in the Great Plains region (South Dakota, Nebraska, Wyoming, Colorado, Kansas, Oklahoma, New Mexico and Texas) and supports more than a quarter of all irrigated farmland in the United States. It is at present suffering severe depletion and increasing groundwater pollution, which is a major cause of concern.

A number of small towns have been built in remote and arid parts of Arabia, Australia and China on sites rich in minerals, petroleum and natural gas. All of these settlements use substantial quantities of water for industrial and domestic purposes. Much of this water comes from fossil groundwater – a non-renewable resource. In some towns, such as the gold-mining town of Kalgoorlie in western Australia, water needs to be brought in from much further away by pipeline, a process likely to accelerate in the future as other mining towns are built in desert areas.

4.3 Adaptations of plants to life in deserts

Desert plants have adapted to the low and erratic rainfall regime in a variety of ways, but the two primary needs are to minimise unnecessary water loss through evaporation while retaining or being able to acquire enough water to allow photosynthesis to proceed efficiently (Evenari et al., 1971; Ozenda, 1977; Evenari et al., 1985; Evenari et al., 1986; Gunin et al., 1999).

The *succulents* retain water in their roots, stems and leaves, and they include the *Euphorbias* of the deserts of Eurasia and the cacti of the American deserts (Figure 4.3). Despite its name, the Barbary Fig (or prickly pear) is not native to north-west Africa

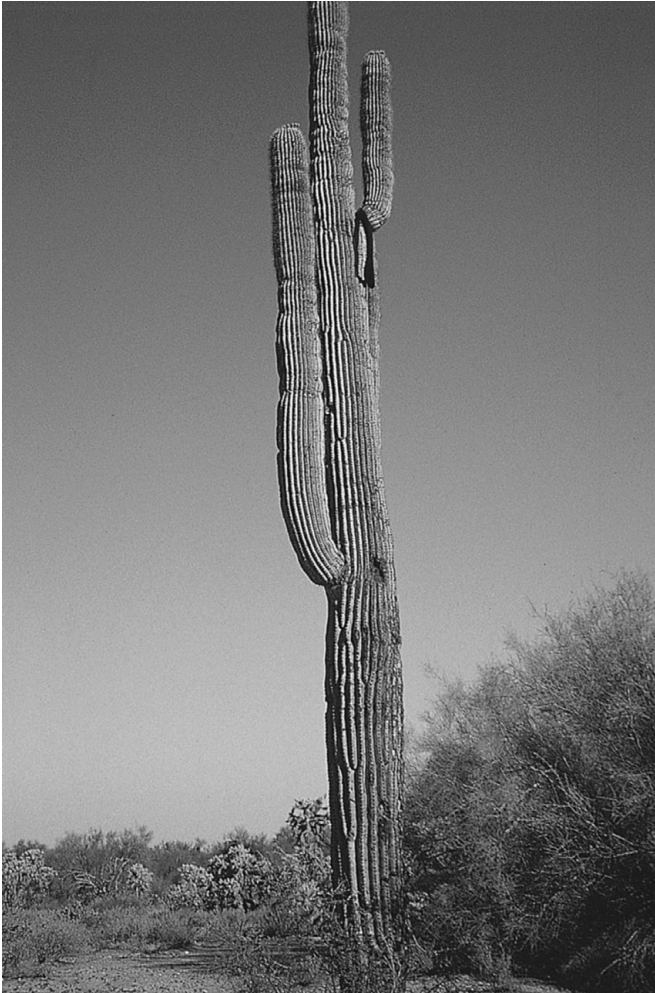


Figure 4.3. Saguaro desert cactus (*Carnegiea gigantea*), near Tucson, Arizona.

but was probably imported originally from Mexico. The mountain summits of the Red Sea Hills are scattered with spectacular *Euphorbia candelabra* trees and derive some of their water from the mists that sweep up in winter from the Red Sea, just as the succulents of the Namib Desert benefit from more than 200 days of coastal fogs a year.

One group of succulents able to store water and tolerate high soil salinities is the desert saltbush (*Atriplex*) of the Australian outback. In fact, during a time of successive droughts that had driven many South Australian wheat farmers in the semi-arid zone bankrupt during the 1860s, the surveyor-general of South Australia, George Goyder, covered 30,000 km (approximately 20,000 miles) on horseback and in 1866 used the southern limit of saltbush and what he had observed about the soils to define what

is now widely known as Goyder's Line. North of this limit rain-fed cultivation was proscribed, but light sheep and cattle grazing was encouraged. The line coincides roughly with the 250 mm (10 inch) rainfall isohyet. His recommendation was ignored but renewed droughts in the 1870s brought home the wisdom of his advice, reflected in many abandoned stone homesteads still very evident today. Other salt-tolerant plants (or *halophytes*) are the Tamarix trees that can excrete salt from their leaves.

Succulents are very rare in the Sahara despite its mixed heritage of plants derived over time from a variety of phytogeographic regions, including the Irano-Turanian, Mediterranean, Saharo-Arabian and tropical African regions. Plants of the Saharo-Arabian region are predominantly *xerophytes*, that is, plants able to resist drought. Xerophytes fall into three broad groups: plants that can cope with desiccation, plants that stay inactive during the dry season and plants that remain active during the dry season.

Lichens are some of the more obvious organisms in deserts that are able to endure desiccation, regenerate rapidly on contact with water, grow anew and, once the moisture supply is used up, dry up and remain dormant until the next supply of rain or dew. Lichens are unusual in that they are not one organism but two – a fungus and an alga – that function together symbiotically. Xerophytic algae also form crusts on the surface of dunes and rocks, and some can remain viable even after several years of desiccation. Very few higher plants possess this ability to survive desiccation without enduring severe damage to their tissues.

Among the xerophytes active during the dry season are shrubs and dwarf shrubs, certain trees and bi-seasonal annuals. The shrubs and trees can reduce dry season transpiration losses by shedding part of the plant, including branches. The disconcerting (and sometimes fatal) impact of large branches falling from eucalypt trees during hot dry spells is a prime example of this process in action. Other adaptations include the replacement of large leaves by spines or needles, the latter sometimes jointed and able to be shed in segments, the development of a waxy cuticle on the leaf surface and the ability to modify the position of the stomata below the epidermal surface. Another form of adaptation involves the root systems of desert plants. Depending on the habitat, the roots may be shallow and of considerable lateral extent, tapping near-surface water in the shallow wetting zone following light rain, or they may extend to depths of many metres to tap the local water-table. In some instances the plants have adopted both strategies, with a combination of deep roots and a considerable lateral spread at shallow depth.

Although devoid of surface water, desert dunes are often reliable repositories of shallow groundwater, because capillary rise of stored water is precluded by the relatively coarse particle size of sand grains, so that water loss from evaporation is minimal. Where sand dunes or sand sheets occur sporadically on alluvial clay plains in semi-arid areas, trees and many grasses will grow preferentially on the sandy soils and avoid the clays, where the soil water is held under tension and is therefore less

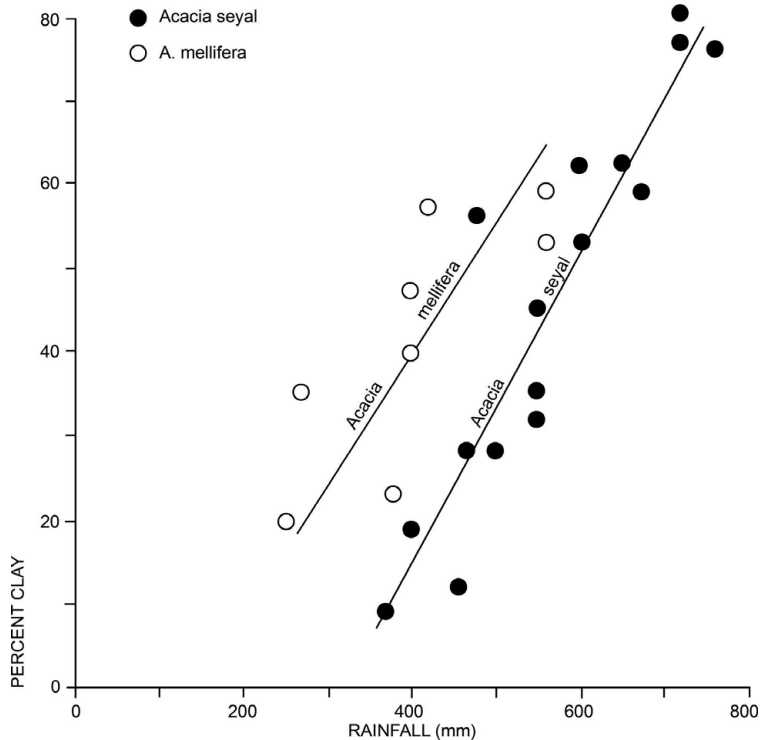


Figure 4.4. Influence of rainfall and soil texture on two acacia species, Sudan. (After Smith, 1949, and Williams et al., 1982.)

available for plant growth than the water within sandy soils. The leafless saxaul tree (*Haloxylon persicum*) is native to the steppes and deserts of southern Russia and central Asia and may be found on dunes extending from the southern Negev Desert to the sand deserts of Inner Mongolia, where it provides shade and fodder for the Bactrian camels and occasional timber and fuel for the Mongolian herders.

In a now classic investigation, Smith (1949) studied the distribution of tree species in Sudan in relation to rainfall and soil texture along a south to north transect from wet to dry and found that a given species of *Acacia* growing on sand needs only two-thirds of the annual rainfall required by the same species when growing on clay (Figure 4.4). Smith was careful to select level sites that neither received water from run-off nor shed water to other sites.

However, the influence of rainfall and soil texture on plants in other semi-arid and arid environments is often more complex than Smith (1949) had inferred more than sixty years ago. Walter (1972) formulated the ‘two-layer’ hypothesis of desert ecosystems (Ogle and Reynolds, 2004). Walter noted that in savanna ecosystems, woody and herbaceous plants could coexist because they drew their water from different layers in the soil. In three influential accounts of desert ecosystems,

Noy-Meir (1973a; 1973b; 1974) described what he termed the ‘inverse soil texture effect’ on plant production in semi-arid grasslands and formulated the ‘pulse-reserve-response’ concept for deserts in North America and elsewhere to describe the influence of pulses of rainfall on soil water and associated plant responses. The inverse soil texture effect simply means that in wet environments, fine-textured soils such as clays and clay loams will support a denser plant biomass than plants growing under a similar wet climate on coarse textured soils such as sands and loamy sands, but as the rainfall diminishes a threshold is reached in which sandy soils support a greater plant biomass than fine-textured clay soils. Later workers have confirmed, refined or modified these two cardinal principles of desert plant ecology (Ogle and Reynolds, 2004). The ‘Westoby-Bridges pulse-response hypothesis’, as it was termed by Noy-Meir (1973b) because it was based on their unpublished data, was identified as a key tenet of desert ecology. In essence, plants (and animals) adapt to erratic rainfall by only responding to certain high rainfall events, which stimulate growth and reproduction and allow the organisms to establish reserves to tide them over in times of reduced rainfall (Ogle and Reynolds, 2004).

Liang et al. (1999) analysed the biomass dynamics and water use efficiencies of five plant communities in the short grass steppe of Colorado, finding more efficient use of water in plants growing on coarse-textured soils, thereby confirming the importance of the inverse texture effect. However, the rainfall threshold, or crossover point, is not everywhere the same and varies with elevation and mean temperature during the growing season, as Epstein et al. (1997) discovered in their studies of the influence of temperature and soil texture on above-ground primary productivity in grassland ecosystems of the semi-arid to arid Great Plains region of the United States. They noted that the crossover point to the inverse texture effect only began when the mean annual rainfall amounted to 800 mm, or far higher than expected from observations elsewhere, and they pointed out that because grassland ecosystems coincided with mean annual precipitation amounts between 250 and 1,000 mm/year, their results showed the need to factor in ambient temperature (itself controlled in part by elevation), as well as rainfall.

Belsky (1990) had earlier investigated the proportions of trees to grassland in East African savannas in relation to the relative influence of variations in rainfall, temperature and soil texture on the availability of soil moisture and soil nutrients during the growing season, concluding that many of the grasslands were ‘edaphic grasslands’, or grasslands determined by soil type. One may note here that the volcanic soils of East Africa differ greatly from those developed on sedimentary or metamorphic rocks in other desert regions, and they display considerable geochemical variation, with some more prone to set hard and impede seedling germination, creating what can be called ‘edaphic drought’, so that soil structure or soil aggregate stability is another important variable influencing plant growth in dry areas.

Reynolds et al. (2004) studied the interactions between plant responses to rainfall variability in the Mojave, Sonoran and Chihuahuan deserts of North America,

concluding that the pulse-reserve-response concept needed some modification to take into account differences in nutrient availability, variations in antecedent soil moisture, and differences in the composition and cover of different plant functional types.

The ability of certain desert grass seeds to remain viable for many years between rains in extreme deserts, such as the eastern Sahara, has given rise to the phenomenon of *gizu* grazing, in which the nomads of northern Sudan and southern Libya will take their flocks deep into the desert during times when unusually heavy rain has led to widespread plant germination and growth. In an original use of the dead remains of such grasses, Haynes (1989) was able to calculate the rate of advance of a barchan dune in the eastern Sahara from the known historic records of the rare rainstorms that gave rise to the *gizu* grasses during the last century.

Fires lit by lightning during convectional storms at the start of the rainy season are common in both tropical savanna and Mediterranean environments. As a consequence, many plants in those regions have adopted a variety of ingenious survival strategies, including thick insulating bark in the case of Mediterranean oaks to rapid shedding of burnt bark by eucalypts. Certain tropical grasses, such as ‘spear grass’ (*Heteropogon contortus*) in northern Australia, seem to require the passage of grass-fires to ensure more efficient seed dispersal. However, other types of vegetation are highly sensitive to fire, including the remnant patches of monsoon rainforest preserved in isolated localities in the monsoonal tropics of northern Australia. The arrival of humans into these regions altered the fire regime and often caused considerable damage to trees during the flowering season. Over time, humans also adapted and modified their burning practices accordingly (Haynes, 1991). But the prairies of North America (and their dependent buffaloes) never survived the burning practices of the European immigrants.

4.4 Adaptations of animals to life in deserts

The physiological and behavioural adaptations of animals living in deserts are designed to avoid water loss and minimise heat stress (Evenari et al., 1971; Newby, 1984). Two other stress factors prevalent in high-altitude deserts, such as the arid high plains or Altiplano of Bolivia and Peru, and the arid uplands of Mongolia and Tibet are extreme seasonal cold and limited supplies of oxygen at high elevations. Extreme cold is also a feature of hot deserts during winter nights, when temperatures can drop well below freezing point. Large mammals minimise heat stress in several ways. The woollen coats of the llamas, guanacos and vicuñas of the Atacama and Patagonian deserts offer insulation against both hot and cold, as do the woollen coats of sheep, goats and camels and the hairy coats of yaks. It is no coincidence that the ancestors of these ruminants were among the earliest animals to be domesticated by the first desert pastoralists. In addition, all of these animals are able to endure higher ranges of body temperature than humans can without suffering undue heat stress.

Other physiological adaptations are the ability to drink large amounts of water when needed and the ability to excrete solids and liquids in highly concentrated form with minimal loss of water. Dingoes in the Australian desert can also obtain water from solids excreted by their puppies and by other animals. The most obvious behavioural adaptation seen in marsupials, rodents and carnivores is to remain dormant in shady places by day and to hunt or seek food by night. The big cats of the Africa savannas are nocturnal hunters, as are the jackals and hyenas, although they are capable of daytime activity if required. Smaller animals, including scorpions, lizards and beetles, will seek refuge in burrows, where the relative humidity is higher than the ground outside during the day and the temperature much cooler. Desert snakes, including the various desert vipers like the African saw-scaled viper *Echis carinatus*, can burrow rapidly into sand and remain almost invisible by day. Snakes and other reptiles can lower their body temperature and metabolic rates during times of cold and increase them gradually on contact with warm surface rocks or soils.

4.5 Adaptations of humans to life in deserts

Human societies that have lived in desert areas for many hundreds of years have practised four main lifestyles: hunting and gathering; pastoralism; rain-fed cultivation; and irrigated farming. For the sake of clarity, these lifestyles are considered separately, but they should not be seen as mutually exclusive, given that pastoral nomads may practise some cultivation, and sedentary farmers may resort to hunting.

The hunter-gatherer lifestyle of the San people of the Kalahari relies on an intimate knowledge of plants and animals across a wide area, as well as an ingenious method of procuring water from sip wells and storing it in ostrich eggshells (Lee and DeVore, 1976). Whether the San people are really what Marshall Sahlins (1968) termed the ‘original affluent society’ is highly debatable. Indeed, as Carmel Schrire remarked, tartly but aptly, the affluence may reside more in the mind of the anthropologist than in the belly of the hunter (pers. comm., Cape Town, June 1979). Until about 100 to 200 years ago, the Aboriginal desert dwellers of Australia relied on an intricate and far flung network of trading and social contacts to enable them to survive in unusually dry years by moving into the lands of their neighbours and using their natural resources (Gould, 1980; Veth, 2005a; Veth, 2005b; Hiscock and O’Connor, 2005). This reliance broke down after the arrival of Europeans on the continent, just as it has done with many other indigenous peoples in the deserts of the Old and New Worlds (Paterson, 2005; Kinahan, 2005).

The hunter-gatherer way of life remained universal until the advent of Neolithic plant and animal domestication, which began more than 11,000 years ago in the semi-arid ‘Fertile Crescent’ region of the Near East (modern Turkey, Syria, Lebanon and Iraq) and then spread into other dry regions like the Nile Valley. Elsewhere, early agriculture developed quite independently, as in several parts of South America,

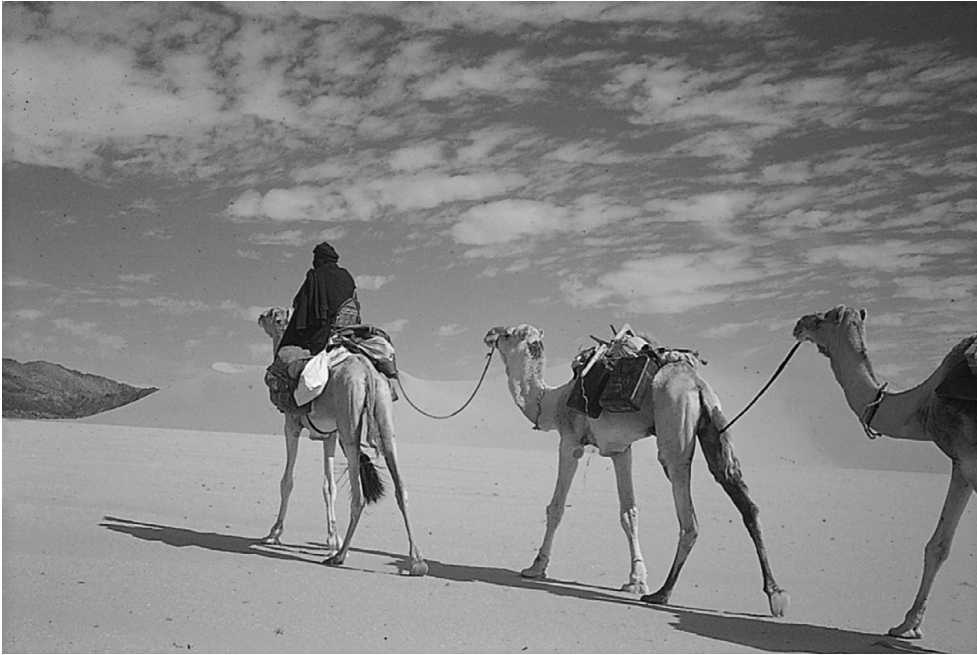


Figure 4.5. Camels crossing sand dunes, south-central Sahara.

Mexico, North America, the Indus Valley and China. To proceed from collecting wild cereal grasses to harvesting domesticated cereals such as wheat, barley, maize, rice, millet or sorghum is simple enough and has only two requirements. These prerequisites are, first, an efficient harvesting tool or sickle and, second, the deliberate collecting, storage and subsequent sowing of mutant plants in which the grains remain on the inflorescence throughout the ripening season, rather than being shed at intervals, as befits the survival strategies of wild cereal grasses (Stemler, 1980). Effective storage from rodents presupposes the use of pottery, which, together with sickles and grinding stones, was already a feature of Mesolithic times before the Neolithic farming ‘revolution’ (see Chapter 17).

The domestication of animals such as goats and sheep, already well-adapted to dry conditions, provided a reliable supply of milk and meat when needed, as well as wool for clothing, blankets and tents. The ancestors of domestic cattle and donkeys were also inured to dry conditions, as were the one- and two-humped camels, often described as the ships of the desert (Figure 4.5). Thus equipped with animals able to survive and even flourish under harsh desert conditions with less than luxuriant plants upon which to graze and browse, the Neolithic inhabitants of both Old and New World deserts appear to have occupied the existing arid lands at a time when they were in fact less arid than they are today. Periodic droughts enabled these desert pastoralists to develop a variety of coping strategies (Figure 4.6), until the inexorable desiccation of



Figure 4.6. Afar mother and daughter with edible lily bulbs collected from a waterhole in the Afar Desert, Ethiopia. These are used in times of drought. The black seeds inside are ground into flour and made into porridge.

the last 5,000 years forced many of them to migrate into areas of more reliable rainfall and pasture. Those who remained, whether by choice or necessity, occupied areas where permanent water was available, either as springs and underground waterholes or at shallow depth in the sandy beds of ephemeral stream channels. Many of these reliable sources of water were in upland areas or in deeply dissected plateaux, but in all cases the availability of adequate food for their animals was a key requirement, often demanding a nomadic lifestyle, leading to an exchange of goods and ideas. The very mobility of pastoralists (Figures 4.7 and 4.8) also gave them a decided advantage in warfare, whether by small-scale raiding parties, such as the Saharan Tibu



Figure 4.7. Stick framework of a portable hut, Jubba Valley, Somalia.



Figure 4.8. Completed hut, Jubba Valley, Somalia.

(Baroin, 2003) and Tuareg tribes (Rodd, 1926; Gast, 2000) until about a century ago or by the detachments of Mongol cavalry led by Genghis Khan (1162–1227) and his successors.

The nomadic existence of desert pastoralists was dictated by the availability of grazing for their flocks, the products of which could be traded for grain, salt and, very much later, sugar, tobacco, matches and other goods. Thus arose a type of symbiotic relationship between sedentary cultivators and highly mobile pastoralists, a relationship that in some instances became institutionalised into rigid social classes, as in the case of the Saharan Tuareg warrior overlords and their Harateen peasant farmer vassals – a pattern still in evidence recently in and around the Aïr Mountains of Niger and the Atakor (Hoggar) massif of southern Algeria (Rodd, 1926; Bernus, 1974). Gast (2000) has documented in great detail the array of plants and other famine foods in historic use by the desert nomads of the central Sahara in times of extreme drought. The knowledge of specific uses for plants is often very localised, sometimes surprisingly so. For example, although many medicinal and culinary uses of the *Boscia senegalensis* tree (family Capparaceae) are widely known (Burkill, 1985), the use of its bark as a water clarifying agent (observed by the writer during the 1974 drought) is apparently only well-known to the Tuareg of the Wadi Azouak region in central Niger.

Where local conditions allow, some nomadic groups have become highly specialized. For example, the itinerant snake catchers of the Thar Desert in Rajasthan, whom the author met in January 1983, move about in small bands of about twenty-five people, with their modest flocks, setting up temporary camps on the dunes at times when various medicinal plants are available for harvesting, including one plant that is dried and kept as an antidote to snake-bites. They travel about 10 km a day with their asses, goats, sheep, dogs and chickens, covering about 1,600 km each year through Rajasthan.

Rain-fed cultivation of cereal grasses such as wheat, barley, maize, sorghum and millet and of a wide variety of fruits and vegetables has been practised in desert areas and their semi-arid margins for many thousands of years. Periodic locust plagues were and are a problem (Figure 4.9). In order to supplement the limited inputs to soil moisture from direct precipitation, several methods have been devised to capture, concentrate and divert run-off to where it is most needed for crop growth. More than 2,000 years ago, Nabatean farmers diverted run-off from the rocky hill slopes of the arid Negev Desert through a series of channels down to the farmland in the valley bottoms (Evenari et al., 1971). Such was their mastery of water harvesting that they were able to build and maintain such cities as Petra – ‘a rose-red city half as old as time’ – in the heart of the desert near Wadi Musa in Jordan (Figure 4.10). Not far from Petra is the spring bearing his name (Ain el Musa) where Moses is believed to have struck water from the rock. The local Bedouin speak of this event quite matter-of-factly, as if it were yesterday.

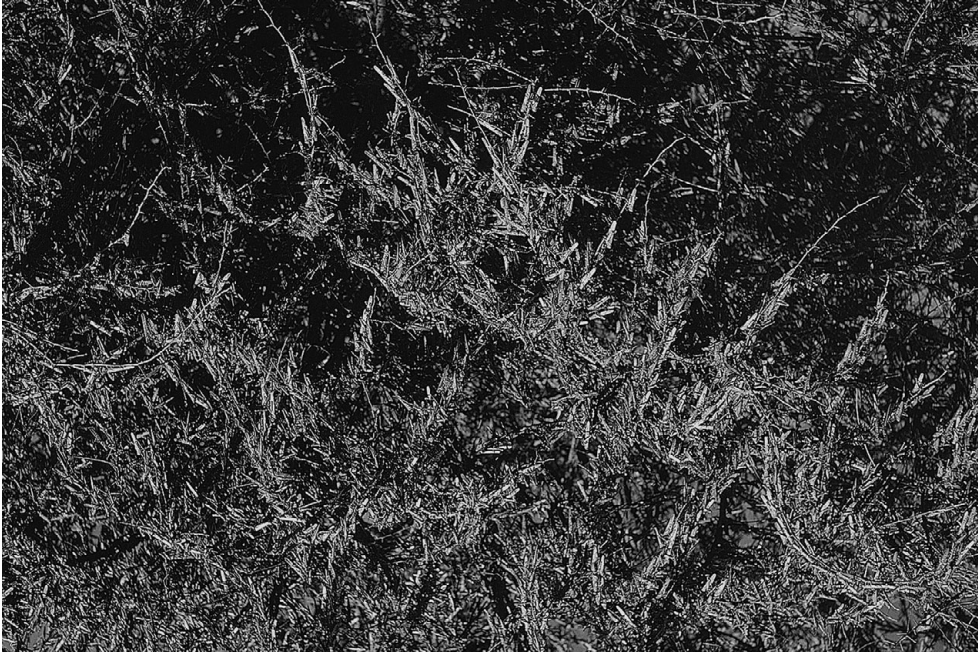


Figure 4.9. Locusts on an acacia tree, Jebel Marra, western Sudan.

The practice of building low earth ramparts, or bunds, around cultivated fields to concentrate water from rain and run-off is still widespread today in the semi-arid alluvial clay plains of Sudan as well as the drier parts of India and Pakistan. These techniques were also used along the valley floors of seasonally flooded rivers like the Indus and the Nile. As skill in impounding water improved, it was a small step from using floodwater to diverting it in canals and using the water for irrigating field crops. An Egyptian mace head shows a pharaoh directing canal digging more than 5,000 years ago, and irrigated farming may have a comparable antiquity in Asia and South America.

One endemic problem with irrigation is the build-up of aquatic plants in slow-moving water in canals, leading to the spread of schistosomiasis, or bilharzia, a parasitic fluke carried early in its life cycle by certain aquatic snails such as *Bulinus* and *Biomphalaria* and passed on as free-swimming larvae into waters frequented by toiling peasant farmers (Williams and Balling, 1996, pp. 137–140). The number of people infected today throughout the tropical world amounts to hundreds of millions, and even the pharaohs suffered from this parasite. Another problem is seepage from canals, bringing salt and waterlogging to low-lying areas adjacent to canals. This has been a major cause of low yields in the Indus Valley and in the great river valleys of central Asia, especially Uzbekistan.

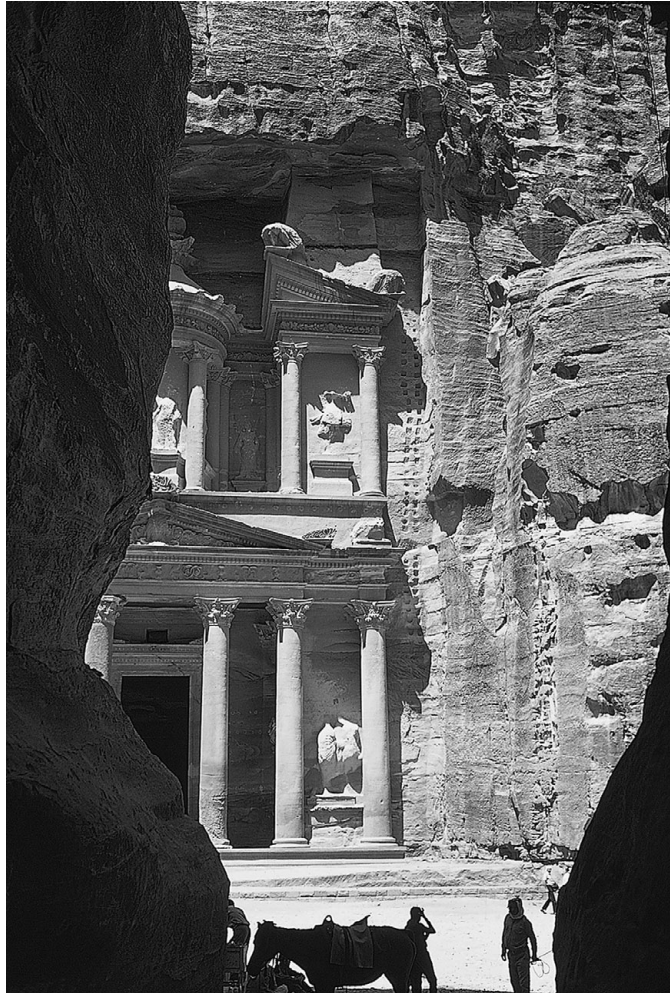


Figure 4.10. Petra, the Nabatean city ‘half as old as time’, Jordan Desert.

Another ingenious method of harvesting water was devised by the Persians more than 2,500 years ago and consists of a chain of wells aligned above a gently sloping underground tunnel which taps into shallow groundwater located within alluvial fans or within the piedmont zone of desert uplands. The water is conducted up to twenty kilometres or more downslope to where it is retrieved at the surface and diverted into a series of channels for domestic use and for irrigation (Figure 4.2). These features go by many names: *karez* in Persian, *qanat* from an early Semitic word (hence ‘canal’ in English) and *foggara* in Arabic. The chain of wells technique was introduced into central Asia and Arabia from Persia more than 2,000 years ago, where a system of well-maintained *karez* provided water for the Persian capital Persepolis, and it was brought to North Africa and Spain by the Arab invaders a thousand years later, and

then from Spain as far as Mexico some 500 years ago (Evenari et al., 1971). The laws governing the use of water from the *foggara* date back more than 2,000 years and are still strictly applied in the remote settlements of the Algerian Sahara (Kobori et al., 1982). The importance of water and its careful distribution in desert settlements in North Africa and the Middle East is reflected in the many hundreds of Arabic words in current use for wells, water containers and methods of collecting water. Nutuhara (1982) recorded Syrian desert villagers using some 350 words relating to water, 70 to water quality, 200 to rain and clouds, 190 to valleys, 140 to pastures, 290 to wells and cisterns, and 110 to ropes as parts of a well. The San people of the Kalahari have a similarly rich vocabulary in regard to hunting and plant food gathering (Tanaka, 1976).

Another method of harvesting water concerns the use of piles of boulders to collect dew. Evenari et al. (1971) were doubtful that humans in the Negev Desert 2,000 years ago could derive much benefit from dew. However, a later reappraisal by Jacqueline Pirenne has shown that dew was an important historic resource for humans and animals in North Africa, the Red Sea coast and islands, Ethiopia, Arabia and the Mediterranean region. She noted that the 'Grand Clapier' in Haute Provence, southern France, which is very dry in the summer, provides 270 litres per hour of water, equivalent to the flow from a kitchen tap (Pirenne, 1977, p. 135). This feature is a ridge of limestone blocks and is 400 m long and 7–15 m wide. The aim of these and other piles of rocks is to allow water vapour to condense on the surface of the stones at night when the temperatures drop down to dew point, that is, 100 per cent relative humidity. The resulting dew then trickles down the face of the rocks and may be captured in small cisterns or may moisten the underlying soil enough to allow some plant growth. In Tigray Province in northern Ethiopia, farmers lead their flocks out before dawn during the dry season to allow them to graze on pastures soaked in dew. Needless to say, other desert creatures benefit from the early morning dew, including birds, lizards and gazelles. In north-west Australia, the irregular mound surfaces of one species of termite trap moisture from dew and were once a resource for Aboriginal people during travel ('walkabout'). The seeds collected by ants and termites and stored in their mounds are still a famine food across the drier parts of Africa, as indeed, are the insects themselves (Gast, 2000).

The long-term adaptations of our prehistoric human ancestors to progressive desiccation and replacement of forest by grassland deserve to be mentioned here. The possible interactions between human evolution and the spread of savanna grasslands in Africa have long provoked interest and controversy. In 1925, Raymond Dart proposed that the expansion of African savanna grasslands in the late Tertiary/Cenozoic may have contributed to the development of an upright posture and bipedal gait in the early hominid that he named *Australopithecus africanus* (Dart, 1925). This 'savanna hypothesis' has been much debated ever since. A recent study by Cerling et al. (2011) of the carbon isotopic ratios (see Chapter 7) measured on soil carbonate nodules

(see [Chapter 15](#)) found in late Cenozoic fossil soils from northern Kenya and the Middle Awash region of the Ethiopian Afar Desert offers qualified support for this hypothesis. They found that the ratio of grassland to tree cover was relatively high during the time that Pliocene hominids were walking upright in these two localities, a topic we consider in more detail in [Chapter 17](#).

Dealing with a much shorter time scale, measured in thousands rather than millions of years, Hiscock and Wallis (2005) and Hiscock and O'Connor (2005) reviewed archaeological and archival studies of historic desert dwellers in the arid south-west of the United States, Patagonia, the Kalahari and Australia and concluded that these modern desert dwellers had undergone a long period of pre-adaptation to living in deserts by having first occupied these areas when they were not as arid as they are today. This argument accords with what is known of the Holocene climatic history of the deserts concerned (see [Chapters 18 to 22](#)) and is reminiscent of Desmond Clark's (1980) suggestion that farming in the Nile Valley, when it did eventually occur, had been facilitated by a long period of pre-adaptation to agriculture through more intensive collecting and grinding of wild cereal grasses in the adjoining, less arid, early Holocene Sahara, a subject discussed in [Chapter 17](#).

Some forms of adaptation to seasonally dry conditions have persisted long after they were no longer necessary – a measure of the reluctance of people long habituated to aridity to abandon well-tried ways. One example will suffice to illustrate this trait. To this day, the cattle herders of the savanna grasslands of the southern Gezira in the central Sudan bring their herds to the White Nile River during the dry season, although abundant water is available in the canal systems situated well away from that river. When the author asked them why they continued this practice, they replied very simply that this was what they had always done.

Over time, many desert communities in widely separated regions of the globe have become committed to conserving both plant and animal resources. For example, the Bishnoi tribe in the Thar Desert of Rajasthan revere all living creatures, especially trees. They will not cut down live trees or branches and will only collect dead wood for fuel. They attach particular importance to the Kajari, or Khejeri, tree (*Prosopis cineraria*, *P. spicigera*), because its edible bark provides relief in times of famine. Elsewhere in semi-arid Rajasthan, it is widely lopped to provide green fodder for the animals, yielding up to 60 kg per mature tree. The branches are then used as fences to protect young animals from predators and finally serve as fuel. In the dry subhumid regions of India, the neem tree (*Azadirachta indica*) is valued and protected, because its leaves contain oil that repels insects, most notably mosquitoes. The desert Aborigines of central Australia had a well-developed systems of totems in which it was forbidden to hunt animals belonging to one's personal totem. In the seasonally wet tropical north of Australia there were and are strong ritual prohibitions against burning in the vicinity of patches of monsoon forest; such forests are very vulnerable to fire and contain abundant valued plant resources. The San hunter-gatherers of

the Kalahari will only take what they require for their immediate needs and show a profound reverence for the animals they hunt.

Among pastoral nomads a similar ethic prevails, but during prolonged droughts differences can arise over access to grazing in areas used by traditional cultivators. Flexible systems of land use, tenure and access to seasonal grazing governed by traditional customs and regulated by locally accepted leaders minimised disputes in the past, but heavy-handed and inflexible central regulations and the breakdown of traditional methods of governance and conflict resolution can exacerbate disputes and lead to widespread conflict and even civil war, as in Darfur.

4.6 Interactions of fire, vegetation and humans

The spread of grasslands at the expense of forests some 7 million years ago was assisted by two main factors: progressive climatic desiccation in the intertropical zone leading to a progressively more seasonal climate with a longer dry season (Zachos et al., 2001) and the physiological ability of grasses to regenerate rapidly after severe fires, in contrast to many species of trees, which are often highly sensitive to being burnt (Kemp, 1981; Singh et al., 1981; Martin, 2006). Once humans had mastered the use of fire – a process that began about a million years ago – they used it increasingly as a hunting tool, further accelerating the spread of grasslands and the demise of former forests and woodlands (Clark and Harris, 1985; Haynes, 1991). The extinction of the large browsers in Australia and the Americas and the proliferation of grazers were at least in part brought about by human use of fire (Williams et al., 1998, pp. 237–238). However, analysis of charcoal abundance in Australasia in late Quaternary sediments not associated with direct evidence of a human presence indicates that fires were most frequent and severe during times of wetter climate and greater availability of potential fuel rather than during the drier climatic intervals, such as the Last Glacial Maximum in Australia some 20,000 years ago (Mooney et al., 2011). It therefore seems that climate rather than human actions exerted a greater control over the incidence of fire, at least until historic times.

4.7 Conclusion

Adapting to life in deserts involves coping with extremes. Rainfall is sparse and erratic, with long intervals of little or no rainfall punctuated by cloudbursts and torrential downpours. Temperatures range from hot to very hot by day to cold or even freezing by night. High-altitude deserts suffer further from reduced oxygen for breathing as well as extreme cold. Plants, animals and dependent human societies have evolved a variety of coping strategies. For plants, which cannot simply move or go underground by night, the physiological adaptations involve minimising water losses from transpiration and making optimum and opportunistic use of the pulses of higher

precipitation in order to grow, reproduce and store food in the form of starch for harsher times. Animals likewise have evolved ways to use water efficiently, to build up fat when conditions are good and to insulate themselves from extremes of temperature with woolly coats, night time foraging or hunting and occupying cool burrows by day. Humans have also devised ingenious ways to divert and store water from below the surface. Desert dwellers have adopted four primary life styles, of which the hunter-gatherer tradition is the most ancient and a carryover from prehistoric times. Nomadic pastoralism originated during the Neolithic some 10,000 years ago, when certain suitable herd animals were first domesticated. Neolithic farmers also practised simple forms of rain-fed cultivation, many of which are still widely used in the semi-arid world today. Finally, in selected localities where perennial supplies of water could be obtained, either from shallow groundwater or from permanent rivers like the Nile, which flowed through the desert, irrigated agriculture was initiated some 5,000 years ago. A major problem facing many desert populations today is the excessive use of groundwater at a rate faster than the current rate of replenishment, in effect depleting these non-renewable fossil groundwater supplies. Pollution of the aquifers from indiscriminate use of pesticides and fertilisers has exacerbated these problems. However, many societies have adapted well to arid conditions in a manner that shows that sustainable use of our arid lands is indeed possible. We return to this topic in the final chapter ([Chapter 26](#)) of this book.

5

Evolution of desert research

The past does not exist. There are only infinite renderings of it.

Ryszard Kapuscinski (1932–2007)

Travels with Herodotus (2004)

(Trans. Klara Glowczewska, 2007)

Il est si facile et si tentant de croire que ‘la’ clef est découverte, que ‘la’ solution est trouvée, comme si d’ailleurs il ne pouvait y en avoir qu’une, précisément celle que son défenseur préconise, alors que dans les choses de la nature, il arrive qu’une même serrure admette plusieurs clefs, et que, par conséquent, une seule théorie, fut-elle la plus séduisante, ou la plus nouvelle, ne puisse prétendre à représenter davantage qu’une modeste vérité partielle et provisoire.

It is so easy and so tempting to believe that we have found ‘The Key’ or ‘The Solution’. The presumption that there is only one answer to a problem – that advocated by the interested party, runs counter to our experience of the natural world, where several keys may fit the same lock, so that no one theory, however original or attractive, can ever claim to represent more than a very modest and provisional part of reality.

Théodore Monod (1902–2000)

The Sahara and the Nile (1980), foreword, xiv, xvi

5.1 Introduction

Our concern in this chapter is primarily with the *historical* record of desert exploration and scientific research, a record that extends back about five centuries for the deserts of South, Central and North America and less than two centuries for the Australian deserts. The written records for Mesopotamia date back to the *Epic of Gilgamesh* (ca. 2700 BC), those for the Nile Valley in Egypt to more than 5,000 years ago and those for China to at least 4,000 years ago. Easily accessible scientific observations in most of the deserts have a relatively short pedigree of only a few centuries, although

desert exploration was well underway before then, often as a precursor to invasion or military occupation. The story of the *prehistoric* settlement of our deserts is best left to a later chapter in this volume ([Chapter 17](#)), but we touch here on aspects of the prehistoric legacy that are pertinent to present-day concerns over desertification.

Herodotus (ca. 485–425 BC) may justifiably be considered the father of desert studies. In the course of his travels in Egypt, he noted the presence of marine shells on low hills near the delta, commenting that the sea must once have been there before the delta had advanced out to sea. He observed that the pyramids were being attacked by salt weathering and also commented on the customs and possible origins of the desert tribes living in the coastal fringes of northern Libya. However, what intrigued him most was why the Nile floods coincided with the hottest and driest three months in Egypt. He inferred, correctly, that the black clay soils along the Nile Valley had come from the Ethiopian Highlands. Following a boat trip north of the delta during which he examined Nile mud carried out to sea, he speculated with remarkable acumen that the abandoned distributaries in the Nile Delta could have become choked with sediment within 10,000 to 20,000 years. He was intrigued by the fact that in the reign of Moeris less than 900 years before his visit, the whole area below Memphis used to become flooded when the Nile rose by only four metres, in contrast to the rise of eight metres needed for flooding to occur when he was visiting Egypt. He concluded that if the flood-plain continued to build upwards at this rate, there would be a progressive reduction in the area flooded. He obtained his information about the progressive decline in the extent of land flooded by the Nile from records kept by the Egyptian priests, some of whose accounts he accepted while discarding others.

Nearly five centuries after the death of Herodotus, another itinerant historian, Diodorus Siculus ('the Sicilian') provided us with a vivid account of cattle rustlers living in the Red Sea Hills of eastern Sudan more than 2,000 years ago who periodically descended onto the plains, rounded up any stray cattle they could steal and disappeared with their booty into the swampy fastnesses of the Red Sea Hills. There are no swamps in these hills today, but there were then, as shown by the presence of permanent freshwater mollusc and ostracod fossils in alluvial sediments dating to that time (Mawson and Williams, [1984](#)).

At about this time, the Roman emperor Nero dispatched two centurions and a cohort of soldiers with instructions to find the sources of the Nile. They failed in their quest but seem to have reached as far south as two low granite hills (apparently Jebel Ahmed Agha, just east of the White Nile, in latitude 11°N), where they reported their way blocked by impenetrable swamps. The nearest swamps today are some 350 km further south, consistent with the former presence of abundant freshwater sponge spicules in sediments and pots of that age ([Figures 5.1 to 5.3](#)) at a site 10 km east of the present-day village of Esh Shawal, situated 350 km south of Khartoum (Adamson et al., [1987a](#)).

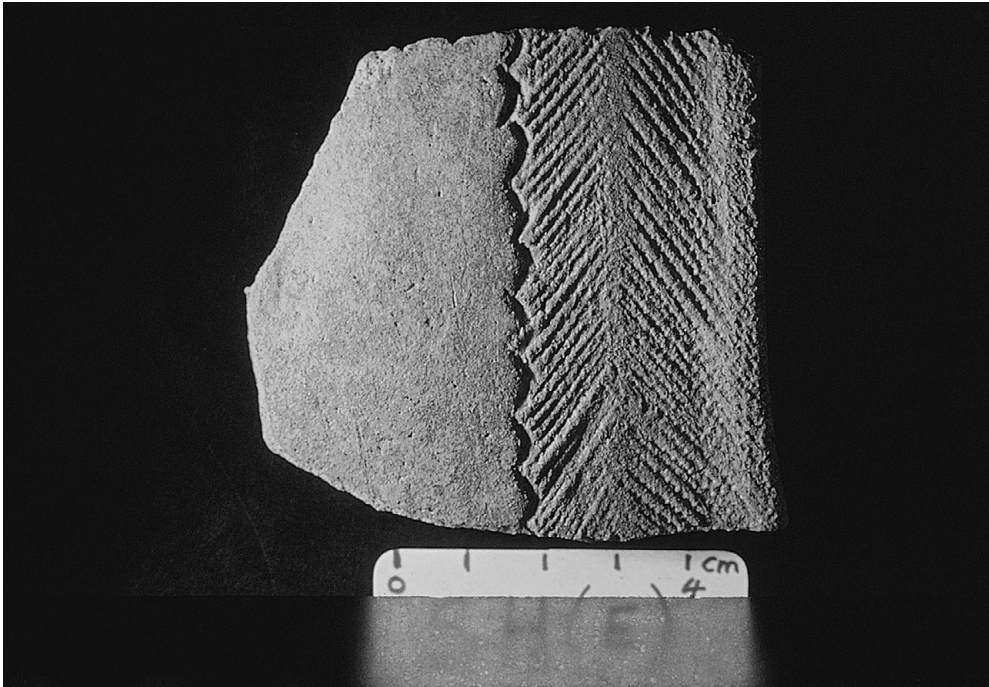


Figure 5.1. Two-thousand-year-old fragment of pottery tempered with the freshwater sponge *Eunapius nitens* (Penny and Racek), Jebel Tomat, lower White Nile Valley, central Sudan. (Photo: Don Adamson.)

The Arab geographer Muhammad Ibn Ibrahim Ibn Battutah (1304–1377) was the greatest desert traveller of all time, covering 120,000 km (75,000 miles) during nearly thirty years (1325–1354) of travel. After his return to Tangier in Morocco, he dictated a lengthy account (*Rihlah*, or *Travels*) of the history, geography and customs of the places he had visited, including Mecca, Persia, Mesopotamia, Arabia, Asia Minor, Bukhara, Afghanistan, India, China, Spain and Timbuktu (see the foreword and commentary by Tim Mackintosh-Smith in the 2012 folio volume of *The Travels of Ibn Battutah* and the abridged text). Another great Arab scholar ‘Abd Ar-Rahman Ibn Khaldun (1332–1406) wrote a monumental history of the Arabs and Berbers (*Kitab al ‘Ibar*) and a later account (the *Maqaddimah*, or *Introduction to History*) describing what he saw as the cyclical progression of nomadic peoples to urban civilization followed by a collapse and a return to less sedentary living. Hourani (2009) provides further details about these two great scholar-travellers in his magisterial *A History of the Arab Peoples*. These accounts of the medieval desert world were familiar to later generations of Arabic-speaking European explorers of Arabia and the Sahara, such as Heinrich Barth (1821–1865), Gerhard Rohlfs (1831–1896), Gustav Nachtigal (1834–1885), Richard Burton (1821–1890) and Charles Montagu Doughty (1843–1926), all of whom published vivid and scholarly accounts of the peoples among whom they

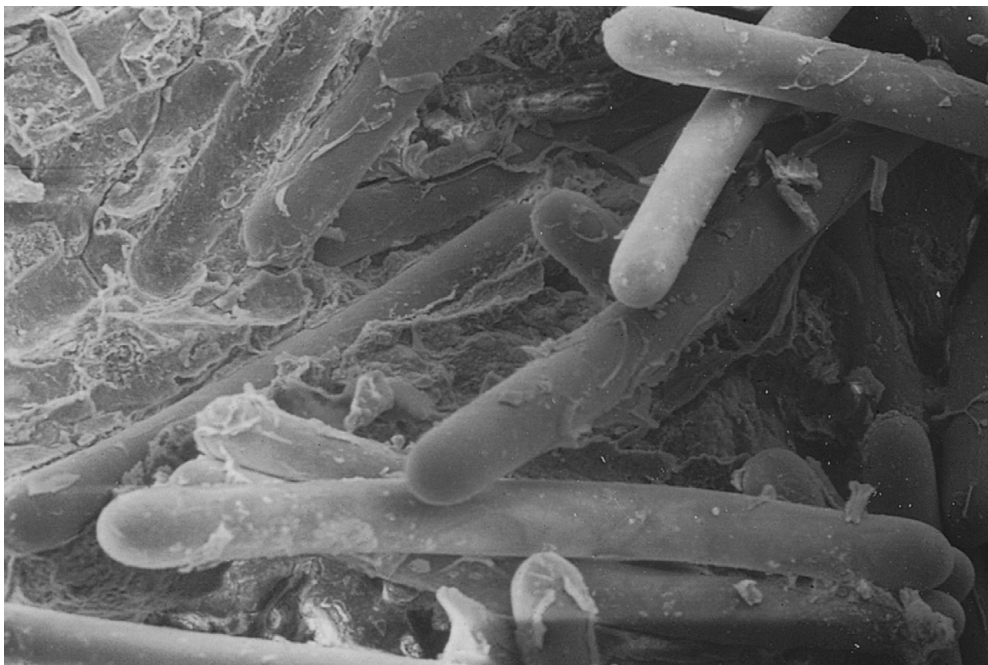


Figure 5.2. Siliceous megascleres in sponge pottery, lower White Nile Valley. (SEM Photo: Don Adamson.)

travelled and of the natural history of the regions they visited. They also received information from the Arab traders they encountered about likely sources of water (Figure 5.4).

From these and other accounts, it is possible to glean useful information about the historical incidence of floods, droughts and famines in this vast region (Nicholson, 1976; Nicholson, 1978; Nicholson, 1980), a topic discussed in Chapter 23. In a remote library in the small desert town of Chinguetti set amidst the dunes of inland Mauritania, there are records of past climatic events written by Arab scholars on vellum that extend back 1,000 years but which have yet to be studied in detail. Many of the older rolls of vellum are abraded. When the author asked why, the custodian said with pride but rather sadly that they had been hurriedly stuffed into saddlebags by his ancestors and taken off into the desert on racing camels for safekeeping during raids from marauding bands over the past 1,000 years.

The maintenance of accurate written records of when certain plants blossomed enabled the Chinese meteorologist Chu Ko-Chen (1973) to use such phenological evidence to reconstruct a temperature history for China covering the last 3,500 years – a remarkable achievement. We are also fortunate in having a detailed annual record of wet and dry years in China for the past five centuries (Anon., 1981). Time series analysis of this outstanding archive has shown that drought years in eastern China were generally coeval with droughts in India and with years of low flow in the Nile

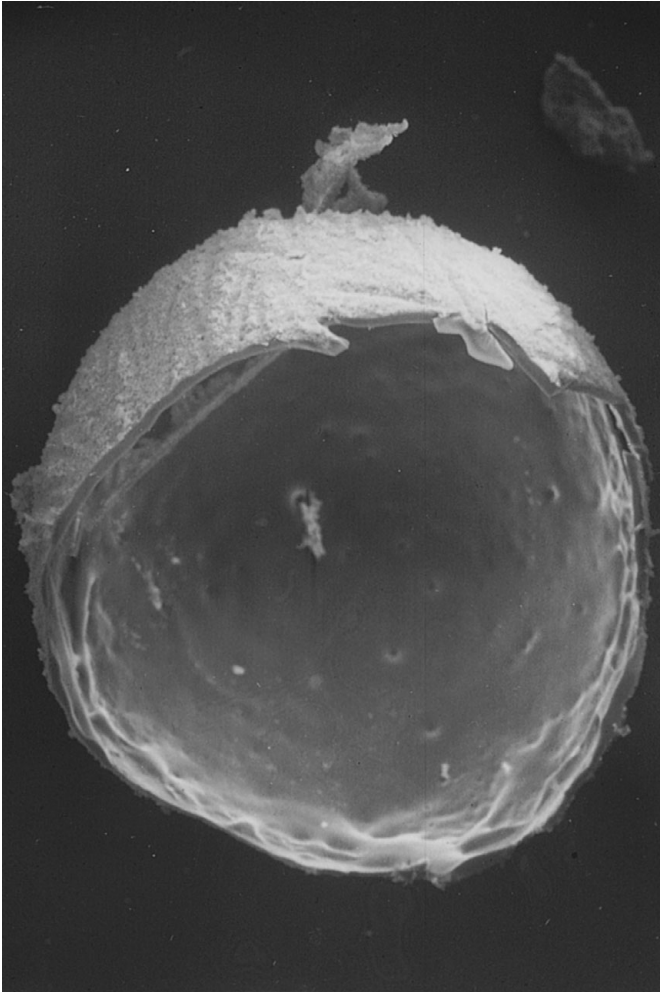


Figure 5.3. Gemmule membrane broken open to show interior, extracted from sponge pottery, Shabona, lower White Nile Valley, Sudan. (SEM Photo: Don Adamson).

during the past few centuries (Whetton et al., 1990; Whetton and Rutherford, 1994; Whetton et al., 1996). We discuss the reasons for this synchronism in [Chapter 23](#).

Lustig (1968) has provided a very detailed and useful annotated bibliography of the geomorphology and surface hydrology of desert environments. It is not our intention here to duplicate his work but simply to highlight some themes and questions raised during the early stages of research on deserts.

5.2 The Sahara and Afar deserts

There are three reasons why it is appropriate to start with the Sahara. First, it is the largest desert in the world, with an area in excess of 8 million km². Second, it has



Figure 5.4. Tamarix mounds indicative of shallow groundwater, Bir Sahara area, Western Desert of Egypt. Wind-blown sand is trapped in the root system of the tamarix tree and develops into a conical mound. These indicators of shallow drinking water were well-known to those who followed the Darb al Arba'in, the notorious forty-day slave route across the desert from Chad to Egypt.

the longest record of prehistoric human occupation of any desert except the Afar. Third, despite its size it is the most intensively studied major desert on the planet. We deal with the Sahara in some detail because it provides a very clear illustration of the successive stages in desert research, from early field observations often by single travellers to detailed multidisciplinary investigations and cognate laboratory studies. Pride of place in these early scientific investigations of the Sahara goes to the French explorers, military patrol officers (*méharistes*) and scientists, to whom we now turn. This is not to discount the outstanding work of British, Italian, German, Spanish, Swiss, Polish, Norwegian and, more recently, American and Canadian geologists and archaeologists, but the French observers paved the way, most notably in the western, central and north-west Sahara.

The themes explored in the next four sections are, of course, not peculiar to arid North Africa but apply equally, with local variations, to every hot and temperate desert on earth. We have begun this account with the Sahara not only because of its great size but also because the scientific discoveries in the Sahara have influenced the search for past climatic fluctuations in other deserts and are mirrored in the scientific accounts that have ensued.

5.2.1 Fossil river systems of the Sahara

Early French observers, like Chudeau (1921), Kilian and Petit-Lagrange (1933), Joleaud (1934), Urvoy (1935; 1937; 1942) and Lambert (1936), commented on the role of previously wetter climates in forming the alluvial terraces so common along the now more or less defunct drainage systems of the central Sahara, such as the Saoura in Algeria or the Azaouak and the Tafassasset in Niger, but were unable to specify when the climate was wetter or how much wetter it was.

The advent of radiocarbon dating in the 1950s (see Chapter 6) revolutionised the study of prehistoric Saharan environments and prehistoric cultures (Alimen et al., 1966). Jean Chavaillon, supervised by Henriette Alimen (Alimen, 1955; Alimen and Chavaillon, 1963), distinguished geologist, prehistorian and founding director of the former CNRS Laboratoire de Géologie du Quaternaire at Meudon-Bellevue near Paris, completed an exploratory study of the depositional history of the alluvial terraces along the Saoura, in which Chavaillon equated fluvial downcutting with increasing discharge and aggradation with a trend towards aridity (Chavaillon, 1964).

Georges Conrad assailed this overly simplified model of river behaviour a few years later (Conrad, 1969). Detailed stratigraphic and paleoecological investigation of the late Pleistocene ‘Saourian terrace’ revealed lignites interbedded with coarse sands fining upwards into shell- and pollen-bearing marls and clays. Clearly, one late Pleistocene depositional cycle encompassed a variety of local and regional environmental fluctuations. Interestingly enough, George Williams’s study of piedmont deposits near Biskra, undertaken independently of Conrad’s (who told me that he was equally unaware of Williams’s work), reached somewhat similar conclusions (Williams, 1970).

The Saharan uplands were logical places in which to study geomorphic changes in the headwaters of once mighty rivers. Pierre Rognon’s detailed and eclectic study of the Hoggar valleys (Rognon, 1967) was shortly followed by comprehensive French, Swiss and German studies of alluvial deposits around Tibesti in which attempts were sometimes made to correlate separate and often undated river terraces with the soon to be outmoded Alpine glacial sequence of Europe (Maley et al., 1970; Messerli, 1972; Winiger, 1972; Jäkel, 1977). Despite later efforts to deduce the likely hydrological conditions under which certain Saharan fluvial deposits had formed (Rognon, 1976a; Rognon, 1976b), it soon became apparent that climatic interpretation of river terraces was fraught with difficulty and all too often riddled with circular reasoning (Williams, 1976a).

5.2.2 Evidence of once widespread lakes in the Sahara and Afar deserts

An awareness of the difficulties involved in correlating and correctly interpreting river terraces, especially when there were no fossils or dateable organic remains,

encouraged French geologists to investigate the former lake deposits that are so widespread along the margins of the Sahara as well as around the central Saharan uplands (Faure et al., 1963). Such investigations also had potential economic value, for certain of the lake evaporites (salt, natron) and diatomites were sufficiently pure and extensive to warrant commercial extraction (Faure, 1963). Well-defined shorelines, formed at a time when now dry or shallow, brackish lakes were deep bodies of fresh water, are sometimes more easily detected in the narrow fault-troughs of the Afar (Fontes et al., 1973) than they are in the vast plains of Mauritania (Chamard, 1973), Chad (Servant et al., 1969) and Niger (Faure et al., 1963; Faure, 1966). Nevertheless, within about a decade, many former strandlines were mapped and dated, fossil hippos and fish were collected (Hugot, 1962; Williams, 1971; Clark et al., 1973; Hugot, 1977) and the paleoecological potential of fossil pollen, spores and diatoms was soon to be exploited. Pioneer studies of small lake basins, *chotts* and *sebkhas* by Roger Coque in Tunisia (Coque, 1962), by Hugues Faure in Niger (Faure, 1966; Faure, 1969) and by numerous others throughout North Africa, often incidentally to their main research (Monod, 1958; Biberson, 1961; Élouard, 1962; Pias, 1971), soon gave way to detailed treatises on major lake basins, such as those of Chad and the Afar. (*Chotts* and their bordering clay or gypsum dunes are similar to the playas of the North American and Mexican deserts and to certain of the more saline pans and lake-lunette complexes in Australia; *sebkhas* are coastal salt flats in arid areas.)

By 1973, Michel Servant had demonstrated that Lake Chad had been high between approximately 40 ka and 23 ka, low from 23 ka to 14 ka and high again after 14–12 ka, with maximum levels around 9 ka. His findings were solidly based on detailed stratigraphic descriptions of deep well sections – a difficult and often dangerous task – backed up by seventy-seven radiocarbon dates – an unusually large number for that time (Servant, 1973). A comparison with the history of the East African lakes (Butzer et al., 1972) strengthened the growing realisation of other researchers that lake level fluctuations in the intertropical zone of Africa during the last 40,000 years had been broadly synchronous from the Atlantic to the Red Sea (Chapter 11).

One important but too little appreciated aspect of the chronology of high and low lake levels just outlined concerns groundwater recharge. Assuming that Saharan lakes were full during periods of sustained high rainfall and low or dry during long dry spells (an assumption supported by the historic fluctuations of Lake Chad), we might expect major transgressions to coincide with phases of maximum aquifer recharge and minimum lake levels to coincide with minimum recharge. Isotopic dating of groundwater in Libya, Sudan and Chad verified this expectation (Sonntag et al., 1980) and showed that many Saharan wells were tapping water dated to 40–25 ka for the deeper aquifers and to 14–6 ka for the shallower aquifers. Sure enough, little or no recharge took place during the long, late Pleistocene dry spell between about 23 ka and 17 ka during which the valleys of mighty rivers like the Senegal (Michel, 1969), Niger and White Nile (Williams and Adamson, 1974) were invaded by wind-blown

sands and active sand dunes were widespread (Mainguet and Canon, 1976) up to 800 km beyond the present confines of the 100–150 mm isohyets, which roughly delineate the boundary between active and fixed dunes today (see Chapter 8).

The diatoms in the lakes of the Afar and the Sahel are sensitive to variations in water chemistry, depth and temperature (Gasse, 1975; Gasse, 1976). The uptake of dissolved silica by diatoms ensures that the siliceous diatom cells, or frustules, accumulate on the bed of the lake as resistant, usually well-preserved fossils (see Chapter 16). Identification to species level may be possible, particularly with recourse to a scanning electron microscope (Gasse, 1974). Diatomaceous lake clays or even pure diatomites may be many metres thick, and because accumulation rates are comparatively slow (0.1–0.5 mm/year is not unusual: Washbourn-Kamau, 1971), a wealth of paleoecological information may be contained within several 20-metre sections sampled at 5 cm or closer intervals. Evolutionary changes in diatoms and the presence of extinct or archaic fossil diatoms within a given lake deposit (Rognon and Gasse, 1973; Servant-Vildary, 1973) allow biostratigraphic correlations to be made with lake deposits much further afield, some of which may lie on or beneath lavas suitable for potassium-argon dating (see Chapter 6).

Gasse (1975) conducted a detailed study of the Plio-Pleistocene evolution and paleolimnology of the lakes of the central Afar Desert. Three types of lakes occur in the Afar: those dependent mostly on surface flow (e.g., Lake Abhe, which is fed by the Awash River); those maintained by a combination of surface and subsurface flow (e.g., Lake Asal); and those sustained by subsurface seepage only (e.g., Lake Afrera). The first type fluctuates mainly in response to variations in the precipitation/evaporation ratio in the elevated headwaters region; the second acts likewise, with a time lag which smooths out minor oscillations; but the third class of lake depends only very indirectly on climatic conditions in the headwaters. Evidence from the Afar lakes shows that during the Holocene, there were three, or perhaps four, periods of high lake level, the fourth apparently coinciding with the latter phases of the ‘Little Ice Age’ of north-west Europe, dated between roughly 1350 and 1850 AD. The Afar late Pleistocene sequence of three high lake-level phases, dated at >40 ka, ≥40 ka to 30 ka and 30 ka to 23 ka, confirmed that major lacustrine transgressions in East Africa were broadly in phase, as was the trans-Sahelian arid interval of 23 ka to 15 ka (Burke et al., 1971; Butzer et al., 1972; Delibrias et al., 1973; Williams, 1975). Dominantly biogenic deposition in Lake Abhe during the interval from ≥40,000 to 30,000 BP raised the question of what the vegetation cover was like in the headwaters of the Awash River at this time, a question that future pollen analysis could answer.

5.2.3 Vegetation history of the Sahara reconstructed from pollen analysis

Almost two decades before paleo-lake studies reached the level of sophistication outlined in the previous section, French scientific explorers were already trying to

reconstruct the vegetation history of the Sahara. Early studies by Pons and Quézel, although soon superseded, showed the pollen bearing potential of even such unlikely samples as calcified crocodile coprolites from the Hoggar (Pons and Quézel, 1957; Pons and Quézel, 1958). During intervals when the winter westerlies brought rain to the piedmont slopes of the Saharan uplands, it was perhaps possible for Mediterranean olive trees to migrate along suitable river valleys into the southern Sahara (Quézel and Martinez, 1958–1959; Quézel and Martinez, 1962; Wickens, 1976a; Wickens, 1976b). *Olea laperrinei*, for instance, grows today on the slopes of Jebel Marra volcanic caldera in western Sudan (Wickens, 1976a; Wickens, 1976b) as well as high on the bouldery, granite slopes of the northern Aïr Mountains of Niger (Quézel, 1962).

Sporadic pollen identifications from ill-dated and often nearly sterile deposits eventually gave way to a more rigorous concern for the ecological requirements of the existing flora, the present-day pollen rain and the need for coring and sampling at more appropriate sites (see Chapter 16). Françoise Beucher's floristic and palynological survey of the Saoura Valley (Beucher, 1971) is an excellent example of the first concern, and Jean Maley's early research into the Holocene history of the Chad flora exemplifies the last two preoccupations (Maley, 1981). French scholars and official grant-giving bodies in and outside of France began to recognise that pollen studies are necessarily long and slow, so the plethora of rapidly published but scientifically dubious studies of the early 1960s later yielded to a few weighty monographs (Bonnefille, 1972; Maley, 1981), together with some more concise overviews of the vegetation history of East and North Africa (Rossignol and Maley, 1969; Livingstone, 1975; Bonnefille, 1976).

5.2.4 Prehistoric occupation of the Sahara: Neolithic grazing and desertification

Geomorphic and paleobotanical studies of the complex array of Saharan Quaternary fluvial, lacustrine and wind-blown deposits indicated that prehistoric settlement of the region took place against a background of alternating arid and less arid conditions. The less arid phases were ones of high lake levels, of soil formation, of fluvial activity, and of the migration of plants, animals and small bands of humans into the now arid desert plains. The abundance of Neolithic occupation sites in now arid and empty parts of the Sahara (Monod, 1958; Maley et al., 1970; Delibrias et al., 1976; Clark, 1980; Clark et al., 2008) should not be taken to imply that overgrazing by domesticated herds of Neolithic cattle, sheep and goats created the desert we know today, for the great sand seas, or ergs, were already in existence by Early Stone Age times (Alimen, 1955), more than 500,000 years ago (Clark, 1975). There were sound reasons why Neolithic pastoralists and their herds preferred the sparse pastures of the lake-studded plains around Tibesti, the Hoggar (Rognon, 1967), the Aïr

(Clark et al., 1973) and Jebel ‘Uweināt to the well-watered savanna woodlands further south, with their ferociously biting *Tabanidae* flies (Wickens, 1982), to say nothing of Trypanosomiasis, Sleeping Sickness and Yellow Fever. However, there seems little doubt that the late Holocene in Ethiopia (Gasse, 1975; Williams et al., 1977) and the Sahara was a time when increasing burning and clearing of land for cultivation in the wetter areas and grazing by domesticated herds of cattle, sheep and goats in the drier areas would have accentuated the biological pressures exerted by climatic desiccation on a flora and fauna adapted to the moister climates of the early Holocene, a topic to which we return in later chapters.

5.3 The deserts of peninsular Arabia and the Levant

Crossing the Rub’ al Khali Desert or Empty Quarter of peninsular Arabia was something generally best avoided, according to the Bedu camel herders who accompanied Bertram Thomas (1931), St. John Philby (1932) and especially Wilfred Thesiger (1946–1950) on their travels in that region. Despite the difficulties, Thesiger in particular brought back some useful survey data and left a vivid account of his experiences (Thesiger, 1959). With the discovery of oil, geological exploration proceeded apace, providing an impetus for detailed studies of dunes (McKee, 1979), desert sediments (Glennie, 1970) and Quaternary landforms indicative of past climatic fluctuations (Al-Sayari and Zötl, 1978). Earlier work had concentrated more on the evidence for a prehistoric human presence in this now waterless region and on excavating the long-abandoned cities in former Mesopotamia, including Uruk in the lower Euphrates Valley, whence came the clay tablets on which was inscribed the *Epic of Gilgamesh* (Sandars, 1972; see Chapter 12).

Three of the world’s great monotheistic religions arose in the drylands east of the Mediterranean – a region that has attracted the interest of pilgrims, biblical scholars and scientific explorers ever since. The 2,000-year-old Nabatean city of Petra enticed archaeologists to the area north of Wadi Rum in the Jordanian desert. Occasional Nabatean grindstones may be seen today near the entrance to Wadi Rum, where springs used to emerge high in the cliffs at the contact between the underlying Precambrian granites and overlying Palaeozoic sandstones. Further north is Wadi Musa and the spring where local tradition indicates that Moses struck water from the rock. In the arid Negev Desert of southern Israel, there are also abundant remains of Nabatean and later Byzantine settlements, as well as prehistoric remains dating back to the Lower Palaeolithic (Evenari et al., 1971; Horowitz, 1979; Ginat et al., 2003; Goren-Inbar et al., 2004; Avni et al., 2006). The evidence of higher lake levels in the Dead Sea Rift drew the attention of nineteenth-century observers already familiar with accounts of much expanded lakes in now arid areas, such as central Asia. In his 1865 report on the geology of the Dead Sea region, Hull was perhaps the first to coin the term *pluvial* to indicate a time when lake basins in now arid areas experienced wetter conditions



Figure 5.5. Mud-brick fort abandoned as a result of climatic desiccation some 2,000 years ago in Xinjiang Province, north-west China.

(Flint, 1971, p. 19). The Dead Sea lakes have been studied in detail ever since (Enzel et al., 2006).

5.4 The deserts of Asia

The first Western knowledge of the deserts of central Asia and north-west India arose from the military campaigns of Alexander the Great in 330 BC, and it was followed by more peaceful interactions involving trade between Rome and India three centuries later and between China and Rome more than 100 years after that (around 100 AD). The opening up of trade between China, India, central Asia and Europe involved establishing trading posts and sporadic military garrisons. A number of these posts were located along the northern and southern borders of the Taklamakan Desert in the now virtually waterless Tarim Basin (Figure 5.5).

The ancient pluvial Lake Lop Nor, located between the Gobi Desert to the east and the Taklamakan Desert to the west, once supported a flourishing garrison settlement at Loulan. From 1899 to 1902, Sven Hedin discovered the beautifully carved wooden remains of this abandoned outpost, together with documents, as did Aurel Stein a few years later, during his travels in 1906–1908. These two explorers brought the story of the former expansion and gradual shrinkage of Lop Nor soon after the third century to

a wider audience (Wood, 2002). The geographer Ellsworth Huntington (1907; 1945) invoked historic droughts following earlier pluvial climates as a primary cause of the periodic migrations of the Mongol nomads into China and Russia. The geological expedition led by Berkey and Morris (1927) into Mongolia also noted evidence of previously wetter climates in this presently arid region.

The German geographer-explorer Baron Ferdinand von Richthofen referred to these historic trade routes collectively as the *Seidenstrasse*, or the *Silk Road*, a name that immediately attracted popular romantic interest (Wood, 2002). Chinese officials kept detailed lists of goods traded and of the peoples involved from the time of the Han dynasty onwards. They also recorded unusual weather phenomena, dust storms and when certain flowering plants blossomed, all of which provide an unrivalled archive of historic climatic fluctuations (Anon., 1981; Chu, 1973; Godley, 2002; and Chapter 23). Richthofen (1877–1885; 1882) also drew scientific attention to the great loess deposits of central China.

Nineteenth-century rivalry between the expanding Russian Empire and the British in India led to clandestine forays by both sides to map the rivers, lakes and mountains in the huge region between central Asia and the Himalayas, including Tibet and Afghanistan (Hopkirk, 2010). The once independent desert cities of Bokhara, Tashkent and Samarkand later fell under Russian control, while the British maintained an uneasy presence in northern Pakistan and Afghanistan. Further east, scholarly scientific explorers such as Sven Hedin (in 1897 and again in 1900–1901) and Nikolai Przhevalsky (from 1870 to 1872) documented the natural history and geography of Asia's largest desert, the Gobi, which covers some 1.3 million km², as well as the Alashan Plateau in what is now Inner Mongolia and the Taklamakan desert. The geological work by Berkey and Morris (1927) provided the first substantive account of Mongolian geology and had a major influence on later accounts of arid zone geomorphology (e.g., Cotton, 1947). The Yale University geographer Ellsworth Huntington carried out sustained fieldwork in central Asia early last century and became well-known for his vigorous espousal of environmental determinism, in which he argued that human actions are determined by environmental and, ultimately, by climatic changes (Huntington, 1907; Huntington, 1945).

In lieu of huge deserts and evidence of vast former pluvial lakes, India provided evidence of a former flourishing urban civilization – the Indus Valley Culture – in the form of the abandoned cities of Harappa and Mohenjo-Daro (Wheeler, 1968; Allchin and Allchin, 1982). These Harappan settlements were often located on the banks of now defunct river systems, prompting speculation about climatic desiccation, river capture, tectonic displacement of drainage and Aryan invasions around 4,000 years ago (Singh, 1971; Singh et al., 1974; Misra, 1983) (see Chapter 12).

Another feature of the Indian landscape is the presence of laterite, a soil or weathering profile depleted of bases and silica and enriched in hydrated oxides of iron and aluminium. First defined by Buchanan (1807, pp. 440–441) as a clay soil on the

uplifted coastal plain of Malabar in India, which hardened on exposure to air, the term *laterite* was later expanded to include such a variety of ferruginous formations that it ceased to have much diagnostic value (Paton and Williams, 1972). The key point about laterite is that it was considered to be diagnostic of hot, wet tropical conditions, so its presence in fossil form in deserts would indicate a previously hot and wet climate. The reality is more complex and is reviewed in [Chapter 15](#).

5.5 The Australian deserts

Although hunter-gatherers have occupied the drier parts of Australia for more than 40,000 years, they left no easily decipherable record; many of the rock paintings and engravings scattered throughout the arid zone are not well-dated and are often hard to interpret (Smith, 2013). However, they had an acutely detailed knowledge of desert landforms, plants and sources of water, all of which they encapsulated in oral traditions of song and dance, as well as the brilliantly imaginative Dreamtime stories designed to make sense of major landscape features in the deserts. Aboriginal guides were also of great assistance to the early European explorers who traversed the unmapped and often waterless interior, which they described in detail in their diaries.

The heroic era of explorations by men like Ludwig Leichhardt (1813–1848?), Edward John Eyre (1815–1901), John M'Douall Stuart (1815–1866) and Ernest Giles (1835–1897) revealed that many of the large inland drainage systems were highly ephemeral in their flow regime and that the fabled inland seas were chimerical, with large salt lakes as harsh and unwelcome relicts of previously wetter conditions.

This initial phase of exploration was soon followed by the expansion of the pastoral frontier, with wool from sheep as a source of great wealth for a few people (Powell, 1976; Powell, 1978). In the semi-arid margins, the native eucalypts were felled and fired with great energy, prompting a prescient observer, the naturalist Paul de Strzelecki (1797–1873), to remonstrate against the accelerated loss of top-soil caused by the loss of a protective cover of plant litter (Strzelecki, 1845). He noted that the burning caused a reduction in soil organic matter and a breakdown in soil structural stability, leading to reduced infiltration of water into the soil, increased run-off from the bare surface and further loss of top-soil. In his *Physical Description of New South Wales and Van Diemen's Land* (1845), Strzelecki concluded that '*the only effective check upon the influence of that denudation is the preservation either of such scanty vegetation as does exist, or, at least, of the woody fibre, which more or less contributes to the fixing and consolidating of the soil*'.

Another observant geologist and naturalist was Charles Darwin, who travelled from Sydney through the Blue Mountains and on to the Bathurst high plains during a time of high temperatures and strong dust storms, which he found very trying. Darwin was highly intrigued by the unusual flora and fauna in the semi-arid landscape, dubbing it a 'second creation'.

A century later, scientific attitudes towards the arid interior of Australia became curiously polarised (Powell, 1988; Powell, 1991). The eminent geographer Griffith Taylor argued with great eloquence that aridity would exercise a crucial control over future settlement of the continent (Taylor, 1949). His earlier maps showed much of the interior labelled as ‘useless’, and he had predicted that by the turn of the twentieth century, Australia would probably have a population of about 20 million people but not more. His opponents ignored the fact that 70 per cent of the continent was arid or semi-arid and in a burst of misguided optimism wrote in glowing terms of ‘Australia unlimited’, with a forecast population of more than 100 million people by 2000 (Powell, 1993; Flannery, 1994). In any event, Taylor was remarkably accurate in his forecast but was compelled by popular opinion to quit his adopted land for many years.

In the years between the two world wars, geologists, soil scientists, hydrologists and natural scientists made steady progress in mapping and describing the resources of the Australian arid zone. After 1945, there was renewed impetus to map the geology, geomorphology, soils and plant cover in the drier parts of inland Australia, both by the different Divisions of the Commonwealth Scientific and Industrial Research Organisation (especially the Divisions of Soils and of Land Research and Regional Survey, later combined into CSIRO Land and Water) and by the former Bureau of Mineral Resources (now Geoscience Australia). There were also strong efforts by the Australian Bureau of Meteorology to develop a comprehensive grid of meteorological stations across the entire continent and to provide farmers and graziers with reliable forecasts of extreme weather conditions. Considerable effort is presently devoted to analysing regional rainfall and temperature trends and to using the now well-established links between sea surface temperatures around Australia and floods and droughts to predict such events in sufficient time for farmers to be able to act on the information and plan ahead intelligently.

The sandy deserts attracted the attention of geologists and biologists and resulted in early studies of the morphology, spacing and orientation of linear dunes in the Simpson Desert and across the continent (Madigan, 1936; Sprigg, 1959; Jennings, 1968; Mabbutt, 1968; Twidale, 1972; Sprigg, 1979). These studies were pursued in greater detail by later workers who sought to establish the age of the dunes and the processes shaping dune development (Wasson, 1984), an interest that persists to this day (Fujioka et al., 2009; Hesse, 2010) and which we consider in detail in [Chapter 8](#). Early studies of wind-blown dust pioneered by Butler and his colleagues in the former CSIRO Soils Division (Butler, 1956; Butler, 1974; Butler, 1982) were concerned with the role of desert dust or loess in soil formation. These studies burgeoned ultimately into joint research between earth scientists from China, led by the late Professor Liu Tungsheng, and a team of Australians who were among the first foreign scientists invited to visit China in 1975, after the isolation of earlier years (Wasson, 1982). The dust storms that irked Darwin have given rise to careful study, especially by McTainsh

(1989), who before settling in Australia had worked in northern Nigeria monitoring the dust flux of the Harmattan (McTainsh, 1980; McTainsh, 1984; McTainsh, 1987). Hesse (1994) investigated long-term trends in dust flux preserved in deep-sea sediment cores from the Tasman Sea and managed to tease out a climatic signal from changes in dust accumulation rates (see [Chapter 9](#)).

The fossil river systems of the Riverine Plain in south-east Australia have been the object of investigation since the time when these fertile alluvial plains were used for irrigated farming more than a century ago, not least because the buried channels, or ‘deep leads’, were often rich in alluvial gold. Schumm (1968; 1969) introduced the concept of ‘river metamorphosis’ to his peers in Australia during his pioneering research into the ‘prior channels’ and ‘ancestral channels’ in the lower Murrumbidgee basin. Bowler (1978a) built on the foundations established by Schumm and earlier workers in his research on the age and hydrologic significance of the different generations of paleochannels still clearly visible on the surface of the Riverine Plain, as discussed in [Chapter 10](#).

The lakes of inland Australia have also provided useful insights into past changes in hydrology and climate (Bowler, 1978b; Bowler, 1981). Lake Eyre is at present a vast salt pan which is dry in most years but receives occasional floods via Coopers Creek and other rivers flowing through the ‘Channel Country’ of south-west Queensland during exceptionally wet years, including 2011 and 2012. Formerly higher lake levels are evident in the Quaternary lake sediments and associated shorelines preserved in favoured localities around the lake (Magee et al., 1995; Magee, 1998; Magee and Miller, 1998; Magee et al., 2004). Other desert lakes also show signs of vastly greater depth and extent than they exhibit today (Bowler, 1981; Bowler, 1998), and insofar as their former shorelines have been reliably dated, they throw some useful light on the climates of the past, as discussed in [Chapter 11](#).

One lake, which prompted a re-appraisal of whether the Last Glacial Maximum (LGM: 21 ± 2 ka) was drier or wetter than today, is Lake George near Canberra. This topic is discussed in detail in [Chapters 11](#) and [12](#), so only a few comments are needed here. Galloway (1965b) mapped the lower limits of late Pleistocene glacial and periglacial deposits (see [Chapter 13](#)) in the semi-arid Snowy Mountains of south-east Australia and deduced from the lower limit of periglacial solifluction deposits that summer temperatures were 8–12°C lower than they are today. The formation of mountain glaciers and small ice caps is controlled by both temperature and precipitation. Low temperatures in winter and seasonal thawing in summer are the primary controls over periglacial frost shattering and downslope movement of debris by solifluction. Galloway (1963; 1965b) noted that according to the evidence then available, Lake George had been a very deep lake (approximately 30 m) at that time. Using a simple water balance model in which monthly evaporation was a direct function of inferred monthly temperature, he concluded that annual precipitation over the lake basin was about two-thirds that of today. He came to a somewhat similar conclusion from his studies in the semi-arid south-west of the United States (Galloway, 1970;

Galloway, 1983), discussed in the next section. The ‘minevaporal’ hypothesis of high lake levels during times of apparently greater aridity has provoked considerable debate over *glacial pluvials* and *glacial aridity*, which we review in [Chapter 12](#).

5.6 The American deserts

5.6.1 North America

Although the North American deserts are quite small when compared to the vast deserts of Africa, Asia and Australia, they are no less inhospitable and harsh environments for plants, animals and humans. The Chihuahuan and Sonoran deserts extend from the south-west United States into northern Mexico. The two other deserts of significance are the Mojave and the Great Basin deserts. In spite of several centuries of European occupation along the eastern third of North America, it was not until the epic 1803–1806 journey by Lewis and Clark at the behest of President Thomas Jefferson that a successful, albeit arduous, overland voyage across the mountainous divide to the Oregon coast removed the psychological barriers to pushing settlement further west (Powell, 1978, pp. 111–118). However, any potential scientific benefits from this expedition were nullified by the century-long delay in publishing the full account of this trip.

The first notable scientific journeys to the far west before the Civil War were those of John Strong Newberry (1822–1892) and of Ferdinand Vandeveer Hayden (1829–1887), both medical doctors and careful amateur observers of the geology of the Colorado Plateau. However, the most significant contributions came after the Civil War. The first of these were the 1869 and 1871–1872 expeditions by the one-armed Civil War veteran of the battle of Shiloh, Major John Wesley Powell (1834–1902), of the entire length of the Grand Canyon (Chorley et al., 1964). Powell’s great contribution here was to demonstrate the efficacy of fluvial erosion in carving out the canyon and to show unequivocally the fluvial origin of the horizontal beds of sedimentary rock that crop out along the sides of the canyon. In one fell swoop, he laid to rest more than a century’s worth of speculation by European geologists as to the work of rivers and the relative efficacy of marine and fluvial erosion in creating extensive level surfaces. Powell’s later work in the Uinta Mountains (Chorley et al., 1964) led to accurate descriptions of geological structures produced by folding. Perhaps his major contribution to geomorphic theory was his formulation of the concept of base level (Baulig, 1950; Chorley et al., 1964). In 1885, the United States Geological Survey was established, with Powell as its founding director. However, the severe drought in the west in 1890 led to stringent budget cuts – an all too familiar story – and geological surveys and associated research suffered.

Grove Karl Gilbert (1843–1918) was one of Powell’s highly capable field assistants, and his subsequent work places him firmly in the front rank of pioneering geologists who worked in the arid west. His *Report on the Geology of the Henry Mountains*

(Gilbert, 1877) deals in detail with the processes involved in fluvial erosion, notably weathering, transportation and corrasion. He noted perceptively that ‘in regions of small rainfall, surface degradation is usually limited by the slow rate of disintegration; while in regions of great rainfall it is limited by the rate of transportation’ (Gilbert, 1877, p. 105; see [Chapter 10](#)). Gilbert returned to this theme in his paper on *The transportation of debris by running water* (Gilbert, 1914), in which he introduced quantitative measures of the relationship between river velocity and the amount and calibre of debris transported. In this sense, he can be seen as the founder of quantitative geomorphology, and his successors built on his early insights (Leopold et al., 1964).

One aspect of this work on processes of erosion concerns the processes of gully erosion and the causes of arroyo incision and sedimentation (Bull, 1964a; Bull, 1964b; Tuan, 1966; Cooke and Reeves, 1976; Graf, 1982; Graf, 1983a; Graf, 1983b). Another aspect of accelerated soil erosion by wind rather than water was demonstrated most forcefully by the suffering caused by the great drought of 1932 in the Great Plains, vividly described in John Steinbeck’s powerful novel *The Grapes of Wrath* (1939) about the exodus from the Oklahoma Dust Bowl. The root cause was the ploughing up of fertile prairie soils to grow wheat without any attempt at soil conservation. Recognition that piecemeal conservation measures were inadequate led to the establishment of the Tennessee Valley Authority, the first attempt at large-scale, integrated catchment management anywhere in the world and a model soon to be followed elsewhere.

Another of Gilbert’s great contributions was his careful mapping of Pleistocene Lake Bonneville, a vast pluvial lake in the Great Basin (Gilbert, 1890). He identified three main shorelines and noted that the highest of these appeared to be contemporary with the last major glacial advance in this region. From this arose the notion of the Pleistocene glacial pluvial climates discussed in [Chapter 12](#). Gilbert also recognised that the weight of water in the lake basin and its subsequent release had contributed to isostatic deformation of the shorelines, although he was not the first to define isostasy – that distinction belongs to Clarence E. Dutton (1889), another of the illustrious pioneering geologists of the arid west (Chorley et al., 1964; Mayo, 1985).

As scientifically trained observers penetrated south and west into the drier parts of North America, their curiosity was aroused by evidence of former human inhabitation in areas now barren and devoid of surface water. The Anasazi desert farmers of Chaco and Mesa Verde in the arid south-west are a case in point. Did they bring about their own demise as a result of the removal of trees and accelerated soil loss, as Jared Diamond (2005) has suggested, or was their demise a result of severe regional droughts at intervals between about 1200 and 1400, as Bryson and Murray (1977) had argued much earlier? Certainly, the evidence from tree rings and from packrat middens is consistent with initially wetter conditions followed by periodic severe droughts.

Another puzzle confronting travellers in the arid south-west arose from the sporadic fossil remains of now extinct large animals (see [Chapter 17](#) for details). Once again,

the question of causes has aroused vigorous and all too often polarised debate (Martin and Wright, 1967; Martin and Klein, 1984).

5.6.2 South America

The Spaniards had crossed the Andes and had successfully established small coastal settlements in the drier parts of Peru and northern Chile from the sixteenth until the eighteenth centuries on the edge of the driest desert on earth, the Atacama, but it was not until the arrival of HMS Beagle in Patagonia in 1832 that scientific enquiry really began. Here, as in so much else, it was Charles Darwin who led the way and posed the relevant questions. Alexander von Humboldt (1769–1859) had earlier climbed the Andes and kept detailed meteorological and natural history observations and was held in great esteem by Darwin, who considered him ‘the greatest scientific traveller who ever lived’, but Humboldt had little involvement with desert research in South America. Darwin spent considerable time during 1832 and 1835 exploring the drier regions of South America (Darwin, 1860). He was intrigued by the discovery of half a fossil skeleton of one of the large extinct mammals in Patagonia; dismissing human predation as a causal factor, he concluded that gradual and almost imperceptible changes in the environment were responsible (op. cit., pp. 171–175). He also found geomorphic evidence that parts of Patagonia and the Cordillera had undergone tectonic uplift (op. cit., pp. 169–171, 314–315). In the Cordillera, he observed abandoned Indian settlements high in the mountains in regions now too arid to support life and speculated that a slight increase in the occasional rains would have allowed irrigation and human life (op. cit., p. 356). In another instance, he concluded that a river in the Peruvian mountains had been diverted as a result of earth movements (op. cit., p. 358). On one occasion in Patagonia, he observed a local soldier striking fire with the flint from a broken arrowhead. Darwin then searched for other arrowheads and concluded that the flint arrowheads in this region were of some antiquity and predated the reintroduction of the horse into South America (op. cit., p. 105).

All of the questions raised by Darwin after his travels in semi-arid South America have been the subject of scientific enquiry ever since. The geographer Isaiah Bowman (1878–1950) led a scientific expedition from Yale University through the Atacama, culminating in his highly readable *Desert Trails of Atacama* (Bowman, 1924). The Atacama is a narrow coastal desert more than 1,000 km long with an area of approximately 100,000 km². Antofagasta in northern Chile has a notional precipitation of 1 mm/year, a somewhat meaningless figure but one indicative of hyper-aridity.

A recurrent question in South America concerns the reaction of the Amazon rainforest to Pleistocene climatic fluctuations (Colinvaux et al., 1996; Haberle and Maslin, 1999; Colinvaux et al., 2000; Colinvaux, 2001; Bush et al., 2009). Both the pollen evidence (Anhuf et al., 2006) and the presence offshore of glacial age arkose (Damuth and Fairbridge, 1970) appear to favour glacial aridity, but not all workers are agreed,

with some arguing for considerable regional variability in rainfall during glacial times (Wang et al., 2004; Bush et al., 2009; Cruz et al., 2009). The pluvial lakes in the arid Bolivian Altiplano and the wetlands in the piedmont areas of northern Chile have long aroused curiosity and have been the subject of a sustained program of drilling, microfossil analysis and dating by U-series and ^{14}C (Sylvestre, 2009; Vimeux et al., 2009).

5.7 Conclusion

One great advantage of working in deserts is that their very aridity has helped preserve the evidence of past climatic events, such as river and lake deposits. Aridity has aided the preservation of former cities such as Uruk in the lower Euphrates Valley, where clay tablets written some 4,700 years ago have provided us with an epic account of the activities of Gilgamesh, king of Uruk. Equally well-preserved are the cities of Harappa and Mohenjo-Daro in the middle Indus and adjacent valleys, inhabited more than 4,000 years ago and thought by some to have been abandoned because of climatic change.

The history of desert research in every arid region follows a very different set of trajectories, depending on local factors and contemporary political, social and economic constraints. Military conquest and the lure of gold controlled initial exploration in South America. Trade determined external links between China, India and Europe some 2,000 years ago. European colonisation of the Sahara soon gave way to scientific exploration, with attention focussed on former rivers and lakes, followed by more detailed studies of fossil pollen and diatoms. In North America, exploration of the Grand Canyon and of the Great Basin lakes provided us with a new set of geological concepts and provided the foundation for quantitative geomorphology. In both South and North America, the presence of prehistoric stone tools and of large extinct animals initiated the ongoing debate as to whether humans or climate change caused the extinctions.

6

Dating desert landforms and sediments

The heavens roared and the earth roared again, daylight failed and darkness fell, lightnings flashed, fire blazed out, the clouds lowered, they rained down death. Then the brightness departed, the fire went out, and all was turned to ashes fallen about us. Let us go down from the mountain and talk this over, and consider what we should do.

Anon. (ca. 2700 BC)
The Epic of Gilgamesh
(Trans. N.K. Sandars, 1972)

6.1 Introduction

In the search for causes of climatic and other environmental changes, it is essential to have confidence that the assumed cause does indeed precede the inferred effect. Needless to say, correlation between two events does not in itself denote causation any more than does mere succession in time. ‘She sneezed and the building collapsed’ illustrates this conundrum. Her sneeze may have caused the collapse, but it is far more likely that the collapse was caused by more fundamental factors and that the timing between sneeze and collapse was entirely coincidental, an example of *post hoc ergo propter hoc* fallacious reasoning (‘after this, therefore because of this’).

In order to establish a logical chain of cause and effect, two things are necessary. One is a physical connection between cause and effect; the other is a *precise* and *accurate* chronology. A date may be very precise, in the sense of it having very small analytical or statistical errors, but if it fails to provide an age for the actual event being dated, perhaps because of sample reworking, then it is not accurate. Accuracy requires that the age obtained using one or more dating methods actually relates to the event being dated, such as the time of deposition of a particular sedimentary unit. It is therefore important to have a clear understanding of the scope and limitations of the more common dating methods that are used to establish when climatic changes

Table 6.1. *Dating methods commonly used in the reconstruction of climatic change in deserts. (Modified from Williams et al., 1998, table A1.)*

Method	Range	Precision
I. Correlation Methods		
Tephrochronology		
Geomagnetic reversals		
Orbital variations		
II. Radioisotope Parent – Stable Daughter		
Potassium-argon	~50 ka to 5 Ga	0.5%
Argon-argon	~10 ka to 5 Ga	0.5%
Radiocarbon (conventional)	0–50 ka	0–6 ka, 60 years; 6–30 ka, 1%; >30 ka, >1%
Radiocarbon (AMS)	0–50 ka	as above
III. Disequilibrium between Parent and Daughter Radioisotopes		
U-series: I, zero initial ^{230}Th	0–250 ka (α -spec) 0–500 ka (TIMS)	1% (α -spec) <0.5% (TIMS)
U-series: II, excess initial ^{230}Th	0–250 ka	10–20%
IV. Trapped Electrons		
Thermoluminescence	0 to 100–500 ka	10%
OSL (optical dating)	0 to 100 ka	7–10%
ESR (electron spin resonance)	0 to 1 Ma	10–20%
V. Cosmogenic Isotopes		
^3He	1 ka to ~3 Ma	10–20%
^{21}Ne	7 ka to ~10 Ma	10–20%
^{10}Be	3 ka to 4 Ma	10–20%
^{26}Al	5 ka to 2 Ma	10–20%
^{36}Cl	5 ka to 1 Ma	10–20%
VI. Chemical Methods		
Amino acid racemisation	0 to 100–500 ka	>15%

occurred in and around the deserts (Table 6.1). Because there are many excellent and comprehensive published accounts of each of the dating methods discussed in this chapter, the interested reader should refer to these to add substance to the brief outlines given in the following sections (Hurford et al., 1986; Aitken, 1990; Geyh and Schleicher, 1990; Walker, 2005).

6.2 Relative and absolute dating

Before the advent of ‘absolute’ dating methods such as potassium-argon or radio-carbon dating, which provide a quantitative calendar age estimate of the sample being dated, earth scientists had to rely on a battery of ‘relative’ dating techniques. These included correlation based on rock or sediment type (*lithostratigraphy*) and on fossils

contained within the rock (*biostratigraphy*). McGowran (2005) provides a scholarly and comprehensive overview of the merits and pitfalls of biostratigraphy, with a particular focus on Cenozoic marine microfossils. Relative ages were obtained according to the standard geological principles of *superposition* and *cross-cutting relations*. In the case of superposition, if bed A overlies bed B, then it is younger, unless the beds have been overturned as a result of folding or faulting. If bed P cuts through bed A, then it is younger than A. These two fundamental principles were already clearly recognised by the Scottish polymath James Hutton (1795) more than two centuries ago, but they did not become widely acknowledged until the Scottish lawyer-geologist Charles Lyell published his three volume *Principles of Geology* some thirty years later (Lyell, 1830–1833). Other relative dating methods include using weathering rinds on individual rocks, lichen patches, soils of different degrees of ‘maturity’, depth and degree of bedrock weathering, or even prehistoric stone tool assemblages, all of which can provide useful preliminary information, but none of which are capable of yielding ages that are both accurate and precise. For this we need to use methods capable of providing an absolute age, some of which are listed in Table 6.1.

Table 6.1 summarises the more common dating methods used in the reconstruction of climatic change in deserts and desert margins, together with their range and precision. They can be grouped into six broad categories. The first category – correlation methods – includes geomagnetic dating, chemical fingerprinting of volcanic ash beds (tephrochronology) and correlation of marine isotope stages inferred from variations in the stable oxygen isotopic composition of marine foraminifera, calibrated against the astronomical orbital time scales. All three methods require independent calibration using absolute dating techniques. The second category of dating methods is where the unstable parent isotope undergoes radioactive decay to produce a stable daughter isotope. Radiocarbon dating, potassium-argon dating and argon-argon dating all fall within this group of widely used dating methods. The third category involves using isotopes in which there is disequilibrium between the parent and the daughter radioisotopes. All forms of uranium-series dating methods are included within this group. The fourth category involves the use of electrons trapped within the lattice structure of certain common minerals (luminescence methods) and tooth enamel or bone apatite (electron spin resonance). Category five comprises certain cosmogenic isotopes, and the best-known method in category six is amino acid racemisation dating. Each method has its own inherent precision and time range, and it is always advisable to use as many independent dating methods as possible. We expand on these points in the next section.

6.3 Atoms, isotopes and radiometric dating

Table 6.1 lists six broad categories of dating techniques that have all been used in the reconstruction of environmental and climatic change in deserts. Correlation methods

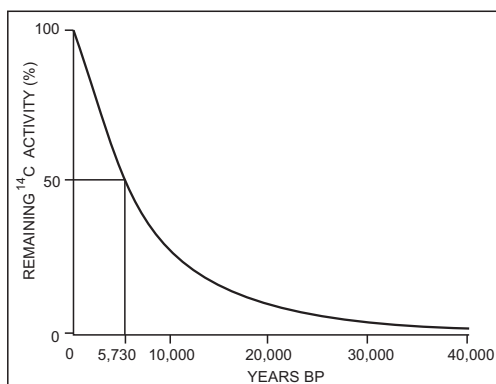


Figure 6.1. Radioactive decay curve showing exponential decrease through time in the relative concentration of a stable radioisotope, in this case radiocarbon (^{14}C) with a half-life of 5,730 years shown on the x-axis and 50 per cent shown on the y-axis. (Modified from Williams et al., 1993, fig. A1.)

(Category I) simply show whether or not a particular rock or sedimentary unit is older than, younger than or of equivalent age to another rock or sedimentary unit, and all should be independently dated to confirm or establish their geological age. The other five categories need some preliminary explanation. For more details, the non-specialist reader might consult standard texts such as Williams et al. (1998, appendix 1) and Walker (2005).

Before embarking on a summary review of dating methods, it will be useful to define certain terms. Although far more complex in reality, in simple terms an *atom* may be considered as consisting of a nucleus of positively charged particles called *protons* and particles with no electrical charge termed *neutrons*. Spinning around the nucleus are tiny, negatively charged particles of minimal mass termed *electrons*. Different *isotopes* of an element, such as carbon, have the same number of protons but different numbers of neutrons. The number of protons + neutrons (i.e., the atomic mass number) is written as a superscript preceding the chemical symbol of the isotope, for example, ^{14}C or ^{18}O . Individual isotopes of an element are called *nuclides*. Some nuclides are stable through time, while others are unstable and emit particles or energy in order to achieve a stable state, in a process termed *radioactive decay*. The atom undergoing radioactive decay is by convention called the *parent nuclide*, and the decay product is called the *daughter nuclide*. The *half-life* is the amount of time needed to achieve equal proportions of parent and daughter nuclide, and it can vary from days to years to millions of years. Radioactive decay follows an exponential pathway, as illustrated in Figure 6.1.

Category II methods, such as radiocarbon dating, are based on the decay of a radioactive element (carbon-14 or ^{14}C) present in the parent sample in very minute amounts. In these cases, the daughter nuclide is stable, in contrast to Category III

methods, in which the daughter nuclide is unstable and is subject to further radioactive decay. Category IV methods, for instance, luminescence dating, are based on the progressive accumulation through time of trapped electrons. Category V techniques, including beryllium-10 dating (^{10}Be), measure the accumulation of cosmogenic nuclides (i.e., nuclides produced by cosmic radiation) in surface and near-surface rocks and sediments. Category VI methods, such as amino acid racemisation, are based on slow chemical changes.

6.4 Correlation methods

6.4.1 Tephrochronology

The term *tephra* refers to ash and other material ejected into the atmosphere during a volcanic eruption, much as described in the Epic of Gilgamesh quoted at the start of this chapter and written some 4,700 years ago. The coarser material is rapidly deposited from the atmosphere, but the finer particles can remain in suspension for weeks, months or even a few years depending on the magnitude of the eruption. After an eruption, some of the ash is transported far away from the volcano to accumulate on land or fall over the ocean, where it settles through the water column and forms a layer of variable thickness on the sea floor. Each volcanic eruption displays a unique geochemical signature, somewhat analogous to human fingerprints, so individual tephra layers can be identified, correlated laterally and dated using some of the methods discussed in this section (Turney and Lowe, 2001). A well-known example concerns the eruption of Santorini volcano on the island of Thera in the Mediterranean Aegean Sea and the burial of the Minoan Bronze Age settlements on the island beneath several metres of volcanic ash. The geochemical tephra fingerprint of this eruption is evident as far away as the Greenland ice cap. Ice core dating of the eruption suggested an age of approximately 1645 BC (Hammer et al., 1987), although ice-layer chronology is not necessarily accurate. Tephra have been widely used in Iceland, Japan, New Zealand and South America for nearly a century to establish a relative chronology of depositional events (Lowe, 2011). The analytical precision involved in finger-printing tephra layers has greatly improved over the past decade, so greater confidence is now attached to what Turney et al. (2004) describe as ‘robust correlation procedures’. Ash layers preserved in lake deposits in arid northern Patagonia have been used to establish a chronology of historic eruptions in the Andes and to correlate lake sediments over a wide area (Daga et al., 2010). In Pliocene Lake Gadeb in the semi-arid southern Ethiopian uplands, ash layers appear to indicate that volcanic eruptions occurred with a frequency of about 5,000 years (Eberz et al., 1988).

One extreme volcanic event that has aroused considerable interest is the eruption of Toba volcano in Sumatra, until recently estimated as having occurred $73,000 \pm 2,000$ years ago (73 ± 2 ka) and now dated more precisely to 73.88 ± 0.32 ka (Storey

et al., 2012). This eruption is considered one of the largest volcanic eruptions of the past 2 million years. It produced at least 2,500–3,000 km³ of dense rock equivalent (DRE) of pyroclastic ejecta, of which at least 800–1,000 km³ was ash (Rose and Chesner, 1987; Chesner et al., 1991; Bühring and Sarnthein, 2000), and the eruption covered peninsular India (located roughly 3,000 km from Toba) in a layer of volcanic ash initially 10–15 cm thick (Williams et al., 2009a), termed the Youngest Toba Tephra, or YTT (Acharyya and Basu, 1993; Shane et al., 1995; Shane et al., 1996; Westgate et al., 1998). By way of comparison, the eruption of Krakatoa in 1883 produced no more than 20 km³ of ejecta, and the 1815 eruption of Tambora produced 30–33 km³ (Foden, 1986; Self et al., 2004). The YTT has been recovered from marine cores in the Bay of Bengal (Ninkovich et al., 1978a; Ninkovich et al., 1978b; Ninkovich, 1979), the Indian Ocean to at least 14° south of the equator, the Arabian Sea and the South China Sea (Pattan et al., 1999; Bühring and Sarnthein, 2000; Song et al., 2000; Liu et al., 2006). Bühring and Sarnthein (2000) noted that because the YTT continues to be found further and further from source, the initial DRE ash volume estimate is likely to be an underestimate, a conclusion endorsed by Williams (2012a) and supported by the very recent discovery of YTT crypto-tephra in a core in Lake Malawi, some 7,300 km from Toba volcano (Chorn, 2012; Lane et al., 2013). (Crypto-tephra are volcanic ash layers invisible to the eye but evident in geochemical analysis.)

Westgate et al. (1998) analysed the major-element composition of the YTT glass shards, as well as their trace element and rare earth element content. They found that the YTT could be clearly distinguished from both the Oldest Toba tuff (OTT) and the Middle Toba tuff (MTT), dated respectively to $840,000 \pm 30,000$ years and $501,000 \pm 5,000$ years ago. The great value of the 74 ka Toba ash outcrops in India is that they provide an *isochronous marker bed* (i.e., one that is of the same age and was laid down at the same time) across the subcontinent, allowing inferred environmental and climatic changes from before and after the eruption to be compared (Williams et al., 2009a; Williams et al., 2010a). For example, the vegetation growing in semi-arid north-central India consisted of forest before the eruption and of grassland or open woodland after it, and the pollen grains preserved in a marine core in the Bay of Bengal also show a reduction in forest pollen in the sediment above the YTT layer in the core (Williams et al., 2009a). However, because the YTT has been to some degree reworked, some workers have questioned its value as an isochronous marker bed (Gatti et al., 2011). In response to this, one can argue that for relatively pure ash (80–90 per cent of the host sediment) to have accumulated in depressions in the landscape as a result of run-off and mass movement, it seems likely that such processes would have occurred quite soon after the deposition of the primary air fall ash, probably no more than a few years later. Distinguishing between primary and secondary ash layers is not always easy. In fact, the recently acquired age of 73.88 ± 0.32 ka for the YTT

came from sanidine crystals in volcanic ash from the Lenggong Valley in Malaysia, located only 350 km from the source (Storey et al., 2012). The very high proportion of pure ash in this alluvial deposit (Gatti et al., 2012) suggests that little time had elapsed between the initial deposition of the primary airfall ash and its subsequent re-deposition in the Lenggong Valley.

6.4.2 Geomagnetic dating

The earth's magnetic polarity has changed periodically by 180°, with the polarity direction sometimes normal, that is, similar to the present, and sometimes reversed. Periods when the magnetic polarity remains stable for a long time (100 ka to 10 Ma) are called *polarity epochs*, with three recognised during the last 3 Ma (Figure 6.2). The current epoch of normal polarity began 0.78 Ma ago (Spell and McDougall, 1992; Brown et al., 1994; Pillans, 2003) and is termed the Brunhes Normal Chron. The preceding polarity epoch (the Matuyama Reversed Chron) lasted from 2.6 to 0.78 million years (Ma) ago, and it encompasses the base of the Pleistocene as presently defined (Gibbard et al., 2010). Shorter intervals of global polarity change (*polarity events*), lasting 10–100 ka, are also apparent in both terrestrial and marine sequences, enabling the record to be used worldwide to establish the age of major environmental changes on land and sea. More short-lived secular geomagnetic fluctuations, termed excursions and often discernible only at a regional scale, have allowed the geomagnetic time scale (Figure 6.2) to be further refined and have been used, for instance, to establish the timing of fluctuations in the level of Holocene and Upper Pleistocene sediments from a lake in Cameroon (Thouveny and Williamson, 1988) and from Pleistocene Lake Bonneville in Utah (Liddicoat and Coe, 1998).

The method is based on the fact that certain rocks and sediments are able to acquire the prevailing direction of the earth's magnetic field, in effect becoming natural magnets (King and Peck, 2001). For example, as lavas cool and start to solidify, the ferromagnetic minerals within them become oriented according to the magnetic field that is prevalent at that time. Likewise, ferromagnetic minerals settling to the ocean floor become similarly oriented, so that long marine cores will show alternating phases of normal and reversed polarity. In the absence of independent dating, this method would only provide relative ages. However, volcanic rocks containing an excellent record of polarity changes spanning the entire Cenozoic have been dated by potassium-argon dating (see Section 6.5.1), so that the geomagnetic record of dated polarity epochs and events provides an imprecise but still useful means of obtaining absolute ages for marine and terrestrial sediments. Examples include the Pliocene hominid-bearing deposits of the Afar Desert (Taïeb, 1974; White et al., 1994; White et al., 2006) and the Miocene fossils of the Siwalik rocks in northern India and Pakistan (Pillans et al., 2005).

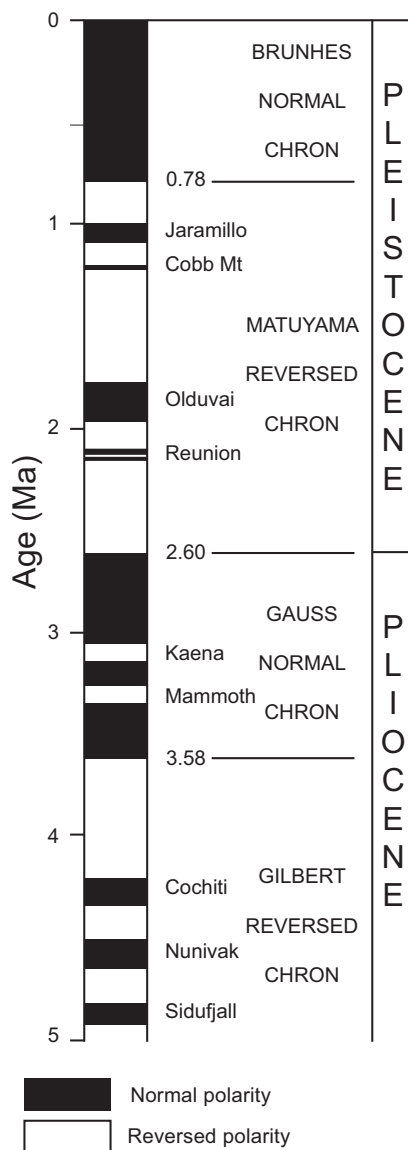


Figure 6.2. Geomagnetic time scale for the last 5 million years. (Modified from Williams et al., 1998, fig. A4, and Walker, 2005, fig. 7.9.)

6.4.3 Dating based on marine isotope stages and orbital (Milankovitch) variations

Initial attempts at devising a chronology based on recurrent expansion and contraction of ice caps during the late Cenozoic, particularly the Quaternary, were doomed from the outset because the terrestrial record of glacial-interglacial cycles is so incomplete (Williams et al., 1998). However, with the advent of stable oxygen isotope analysis

of deep-sea cores, a way out of this impasse was discovered. During glacial cycles, the lighter isotope of oxygen (^{16}O) is preferentially evaporated from the oceans and concentrated in the growing ice sheets, while the oceans become relatively enriched in the heavier isotope (^{18}O). With the interglacial melting of the ice sheets, this lighter ^{16}O isotope is released once more into the oceans. The ratio of ^{16}O to ^{18}O ($\delta^{18}\text{O}$) serves to distinguish foraminifera living under glacial conditions from those living during times of minimum ice volume, according to the expression:

$$\delta^{18}\text{O} = \left(\frac{^{18}\text{O} - ^{16}\text{O}}{^{18}\text{O} + ^{16}\text{O}} \right)_{\text{sample}} - \left(\frac{^{18}\text{O} - ^{16}\text{O}}{^{18}\text{O} + ^{16}\text{O}} \right)_{\text{standard}} \times 1000 \quad (6.1)$$

The $\delta^{18}\text{O}$ units are in parts per thousand (‰, or ‘per mil’). The standard used for foraminifera is PDB, a Cretaceous belemnite from the Pee Dee Formation in North Carolina. (*Foraminifera* are single-celled, mostly marine, planktonic animals with a chalky shell.)

A pair of *marine isotope stages* thus brackets each glacial-interglacial cycle. These stages are numbered from the most recent (MIS 1) backwards. Even numbers denote glacial stages; uneven numbers denote interglacial stages. Analysis of the oxygen isotopic composition of marine microfossils, especially *foraminifera*, has revealed eight glacial-interglacial cycles during the last 800 ka and more than fifty in the last 2.5 Ma (Shackleton et al., 1990; Walker, 2005). The earlier cycles were of shorter magnitude and duration than were those of the last 800 ka, as discussed in [Chapter 3](#).

MIS records correspond closely with the earth’s orbital (or Milankovitch) cycles, and they can therefore be calibrated (‘orbitally tuned’) with respect to the latter. The orbital cycles reflect regular fluctuations in the distance of the earth from the sun and the tilt of the earth’s axis (see [Chapter 3](#)). For present purposes, all we need note is that the three main orbital cycles have a duration of approximately 100,000, 41,000 and 23,000 to 19,000 years, and act as the pacemakers of the ice ages (Hays et al., 1976; Imbrie and Imbrie, 1979). In many instances, the record of environmental changes preserved in marine cores off the coast of desert regions complements the fragmentary terrestrial record and provides a more continuous archive of the climatic changes experienced in deserts and their margins.

6.5 Radiometric dating: radioisotope parent – stable daughter

6.5.1 Potassium-argon and argon-argon dating of volcanic rocks

Potassium is common in many minerals in igneous rocks, particularly feldspars. There are three naturally occurring isotopes of potassium, of which ^{39}K is the most abundant, followed by ^{41}K , with ^{40}K present in trace amounts (0.00118 per cent). The potassium-argon ($^{40}\text{K}/^{40}\text{Ar}$) dating method was developed in the 1960s and involves measuring the ratio of radioactive ^{40}K to the relatively inert gas ^{40}Ar which is a daughter product in its radioactive decay chain, both natural isotopes. Because ^{40}K has a very long

half-life (1,250 Ma), this method is not especially useful for rocks younger than about 50 ka. As a result of the very large error terms involved in dating young rocks using $^{40}\text{K}/^{40}\text{Ar}$ dating, efforts were made to develop a more precise method based on the ratio between two argon isotopes (McDougall and Harrison, 1999). The isotopes in question are ^{40}Ar and ^{39}Ar . ^{39}Ar is produced in the laboratory by irradiating the sample to be dated with fast neutrons in order to convert ^{39}K to ^{39}Ar . The $^{40}\text{Ar}/^{39}\text{Ar}$ method is more precise for younger rocks because of the shorter half-lives involved, allowing samples as young as 10 ka to be dated (Table 6.1). This method has recently been used to obtain a very precise age of 73.88 ± 0.32 ka (1σ) for the super-eruption of the youngest tephra from Toba volcano, known as the YTT (Storey et al., 2012).

6.5.2 Radiocarbon dating of organic and inorganic carbon

Radiocarbon dating is the method most widely used to date late Quaternary marine and terrestrial sediments. Willard F. Libby invented the method (Arnold and Libby, 1949; Libby, 1955), for which he received the Nobel Prize for chemistry in 1960. Like many outstanding scientific discoveries, this one arose quite by accident. In 1947, Libby and his colleagues had collected samples of methane gas produced by Baltimore's Patapsco Sewage Plant and found that it contained trace amounts of radioactive carbon (^{14}C), showing that living organisms harboured this isotope (Balter, 2006). Libby (1973, p. 7) described radiocarbon dating succinctly as 'a measurement of the age of dead matter by comparing the radiocarbon content with that in living matter'.

Radiocarbon is produced in the outer atmosphere by cosmic rays that generate neutrons that then react with the nucleus of stable ^{14}N , detaching a proton, to form the radiocarbon isotope of mass 14 and half-life of $5,568 \pm 30$ years (Libby, 1955). In fact, the half-life is more accurately given as $5,730 \pm 40$ (Godwin, 1962), but for convenience, the original Libby half-life estimate is still used by all radiocarbon laboratories. Because the mean life of any one radiocarbon atom is approximately 8,300 years, there is ample time for its mixing and assimilation in atmosphere, biosphere and ocean. Plants will take in some radiocarbon from the atmosphere during photosynthesis. Marine or aquatic organisms will absorb radiocarbon dissolved in the oceans or in freshwater, and that radiocarbon will become incorporated into their calcareous shells. Soil and lake carbonates, speleothems and tufas (see Chapter 14) likewise absorb radiocarbon dissolved in rain, run-off or groundwater during the time (which may be of quite long duration) in which they are being precipitated. Animals will absorb ^{14}C from the atmosphere as they breathe, and this becomes incorporated into their bones and soft tissues. Once the organisms die, the ^{14}C within the dead organism starts to decay, with half of the ^{14}C converted back to the stable isotope ^{14}N within about 5,730 years and half of what then remains converted to ^{14}N after a

further 5,730 years, until about ten half-lives have elapsed and there are only trace quantities of ^{14}C still remaining (Figure 6.1).

Initial methods of radiocarbon dating were based on measuring the relative proportions of ^{14}C to the stable isotope ^{12}C . The mean isotopic composition of carbon compounds in nature is around 99 per cent ^{12}C , around 1 per cent ^{13}C and 10^{-10} per cent ^{14}C . The stable carbon isotopic ratio δ^{13} is expressed as the deviation in parts per thousand from a standard, according to the expression

$$\delta^{13}\text{C} = (\text{R}/\text{R}_0 - 1) \times 1000\text{‰} \quad (6.2)$$

In this expression, R is the measured ratio of ^{13}C to ^{12}C of the sample, and R_0 is the same ratio for the standard, much as in Equation 6.1. The usual standard in radiocarbon work is once again the Cretaceous carbonate belemnite, *Belemnita americana*, from the Pee Dee formation of South Carolina, known as PDB. Radiocarbon dates are expressed as ages before present (BP), defined as 1950 AD, with a statistical error of one standard deviation, and the laboratory code number. The modern standard has the same count rate as wood grown in 1950 AD (Williams et al., 1998, appendix 1).

Certain qualifications to an otherwise reasonable set of assumptions now need to be made (Faure, 1986; Williams et al., 1993; Williams et al., 1998). First, Suess (1955) found that the ^{14}C activity of twentieth-century wood was nearly 2 per cent lower than that of nineteenth-century wood, as a result of ‘dead’ or non-radioactive carbon emitted into the atmosphere through the burning of coal, oil and gas. This dilution effect is termed the ‘Suess effect’. Second, de Vries (1958) demonstrated that the radiocarbon content of the atmosphere has not been constant but has varied systematically in the past, with ^{14}C activity around 1500 and 1700 AD up to 2 per cent greater than it was in the nineteenth century. This phenomenon is now known as the ‘de Vries effect’. Third, nuclear explosions and increasing use of nuclear reactors and particle accelerators have increased the level of ^{14}C activity in the atmosphere through the input of humanly produced ^{14}C . Fourth, and most significantly, the production of ^{14}C has varied during the late Quaternary as a result of variations in the strength of the earth’s magnetic field, which acts as a shield against cosmic rays. When the earth’s magnetic field is weak, the production of ^{14}C in the outer atmosphere is enhanced, and conversely. Higher values of atmospheric ^{14}C will reduce the age determined by radiocarbon dating, and vice versa.

For all of these reasons, it is evident that the radiocarbon time scale is not a calendar time scale. Radiocarbon ages are sometimes older and sometimes younger than their equivalent calendar ages. Some form of calibration is therefore needed in order to convert radiocarbon years into calendar years. The first successful attempt at calibration came from radiocarbon dating of tree rings, the age of which had already been established from ring counts (Fritts, 1976). This approach has yielded reliable calibration back to about 8,000 years ago, using the long-lived bristlecone pines (*Pinus longaeva*) from the White Mountains of the United States and European oaks

preserved in peat bogs scattered throughout Europe (Pearson et al., 1986). Beyond that time, there have been progressive improvements in extending the calibrated time scale, initially back to 30,000 years (Bard et al., 1990; Guilderson et al., 2005; Blackwell et al., 2006) and later back to 50,000 years using pairs of pristine fossil corals dated very precisely by both ^{14}C and $^{230}\text{Th}/^{234}\text{U}/^{238}\text{U}$ (Fairbanks et al., 2005; Chiu et al., 2005), with the latter method being used to calibrate the former.

Three technical advances made this progress possible. First, the direct counting of individual atoms of ^{14}C using cyclotrons as extremely sensitive mass spectrometers has pushed the *potential* limits of radiocarbon dating back from approximately 40,000 years ago to approximately 100,000 years ago, with 1- to 100-mg samples (Muller, 1977; Stephenson et al., 1979). Second, the use of Accelerator Mass Spectrometry (AMS) and better preparation techniques have reduced the size of the sample needed in radiocarbon dating by a factor of a thousand (Doucas et al., 1978; Muller, 1979; Hedges and Gowlett, 1984). Third, the use of Thermal Ionization Mass Spectrometry (TIMS) has enlarged the range of dating applications that were previously not possible using the less sensitive alpha-counting technique (Chiu et al., 2005; Fairbanks et al., 2005). In establishing their calibration curve, Fairbanks et al. (2005) took great pains to ensure that there had been minimal diagenetic alteration from aragonite to calcite in their coral samples, rejecting any sample with more than 0.2 per cent calcite, in contrast to other workers who used 1 per cent calcite detection limits and calcite sample values of 1 per cent to 5 per cent. The reason for this precaution is that during any chemical alteration from aragonite to calcite, there may have been a loss of radiocarbon.

A major recent breakthrough in calibrating the terrestrial radiocarbon record from 11.2 ka back to 52.8 ka made excellent use of the annually laminated sediments in Lake Suigetsu on the Sea of Japan coast in western Japan (Bronk Ramsey et al., 2012). The age of 52.8 ka is the present limit of the radiocarbon method. The calibration was based on 651 terrestrial radiocarbon dates. One outcome of this work was to show that reservoir ages used to calibrate the Cariaco Basin and north-east Atlantic time scales need some revision and have not been constant through late Quaternary time. It is also worth noting that Lake Suigetsu is an important site for paleoclimatic research in that it provides an annual record of four climate-proxies, namely, winter and summer monsoon intensity and the respective temperatures of the Siberian air mass and the Pacific air mass (Nakagawa et al., 2006). This is because the lake is located north of the monsoon front in winter and south of that front in summer, and so it is highly sensitive to changes in Pacific air mass temperature in summer and Siberian air mass temperature in winter.

Sample contamination by inert or by modern carbon may affect the reliability of the radiocarbon dates (Polach and Golson, 1966; Gillespie, 1982). Figure 6.3 shows the effect of contamination by modern carbon, such as plant roots. The effect is substantial and, unlike the effect of contamination by inert carbon shown in Figure 6.4, is not a

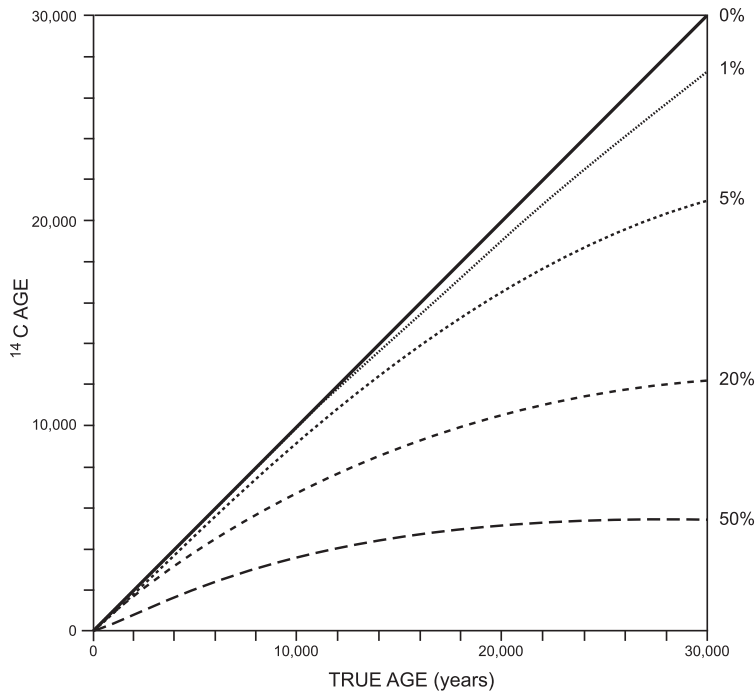


Figure 6.3. Effect of contamination by modern carbon on radiocarbon age. (Graph compiled from data provided in Polach and Golson, 1966.)

linear response. For example, 5 per cent contamination of a sample of true age 30,000 years would appear as a radiocarbon age of 21,000 years. However, sample preparation techniques will in general remove all traces of modern carbon and are today vastly improved on what they were even a decade ago, so this is rarely a significant issue. Far more problematic is the detection of contamination by inert carbon, which is termed the ‘radiocarbon reservoir effect’ (Björck and Wohlfarth, 2001). In most instances, the effect is not very large, with values of only about 400 years not uncommon in many lakes and rivers in the drier regions of Africa and Australia. However, in some instances, the reservoir effect can be significantly large, as in certain early Holocene and late Pleistocene lakes in arid northern Chile, where Geyh et al. (1999) have documented a reservoir effect of at least $-2,000$ years. Rivers flowing to the ocean contain both old (^{14}C -depleted) and young (^{14}C -enriched) terrestrial dissolved organic carbon, and if the young dissolved organic carbon is selectively degraded during transit, the older carbon will preferentially enter the ocean (Raymond and Bauer, 2001). Within the ocean, near-surface reservoir ages may fluctuate through time (Bondevik et al., 2006), so the assumption of a single unvarying reservoir age at any particular locality may be unjustified.

There are particular problems associated with the use of charcoal for radiocarbon dating. We noted earlier the distinction between precision and accuracy in radiocarbon

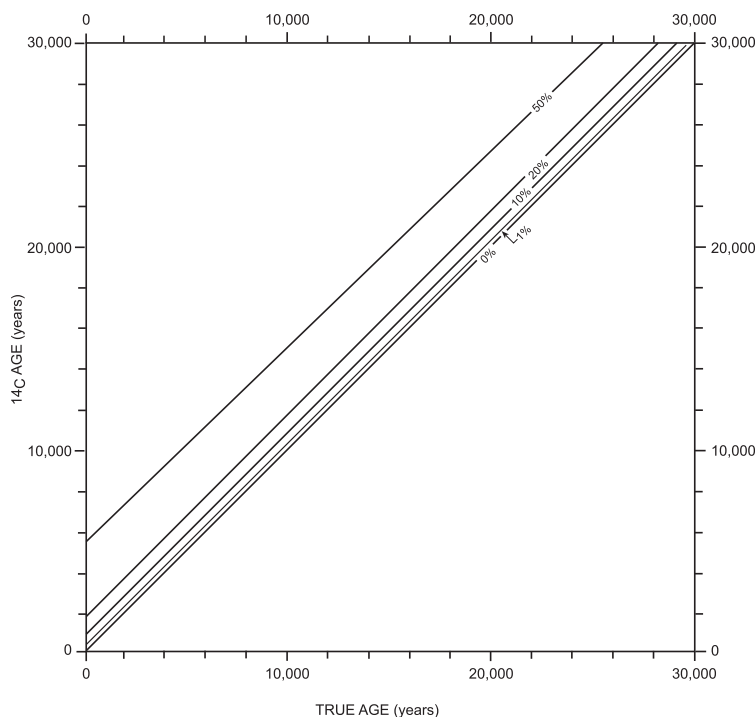


Figure 6.4. Effect of contamination by inert carbon on radiocarbon age. (Graph compiled from data provided in Polach and Golson, 1966.)

dating. Owing to its ability to withstand chemical decay, charcoal may persist for a long time in the landscape in temporary storage sites along the valley sides. Once these stored sources of charcoal are remobilised, perhaps as a result of an extreme rainfall event, they may become incorporated into much younger colluvial or alluvial deposits. A perennial problem in dating river alluvium therefore concerns the possible inheritance of charcoal or shell remains from an earlier depositional cycle. In an elegant study of a small river in the Blue Mountains of eastern Australia, Blong and Gillespie (1978) found that charcoal fragments of varying sizes deposited in channel sand ripples during a single modern storm event ranged in age from 0 to more than 1,000 radiocarbon years, according to fragment size, indicating that previously stored charcoal had been reworked by the flood.

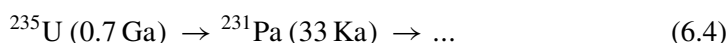
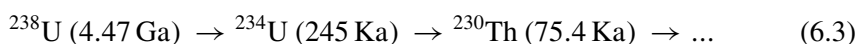
Shells are less likely to survive multiple episodes of transport undamaged, and ones that are broken can therefore be distinguished and not used for dating. A more insidious problem with shells is the possibility that conventional methods of detecting recrystallization from aragonite to calcite using X-ray diffraction and cathodoluminescence may fail to detect what Webb et al. (2007) termed ‘cryptic meteoric diagenesis’ in freshwater snail shells.

Given some of the problems inherent in radiocarbon dating, it is prudent to check the validity of the ages obtained using other, independent methods of dating, such as luminescence, uranium-series or amino acid racemisation (Prescott et al., 2007). In addition, and wherever possible, different types of paired samples should be collected for radiocarbon dating, including charcoal, shells and organic plant detritus in the case of alluvial sediments (Williams et al., 2001; Haberlah et al., 2010a; Haberlah et al., 2010b).

6.6 Radiometric methods using disequilibrium between parent and daughter radioisotope

6.6.1 Uranium-series disequilibrium dating of speleothems, calcretes and corals

With analytical improvements in this method during recent decades leading to greater precision and accuracy of ages obtained, uranium-series disequilibrium dating has been used ever more widely. The method is based on initial decay of the three radioactive isotopes ^{238}U , ^{235}U and ^{232}Th , which have half-lives ranging across the entire spectrum of geological time. The end members of each of the decay series are stable isotopes of lead, but daughter isotopes formed during the decay process are themselves unstable, so that a whole series of decay products are involved at different stages, as shown in Equations 6.3 and 6.4, in which ^{231}Pa is the isotope protactinium-231. (Ga is 10^9 years and ka is 10^3 years.) Samples suitable for dating using this technique include corals, speleothems, evaporates, bones, shells, peat and even weathered alluvium, but not all attempts have yielded reliable results (see Gustavsson and Högberg, 1972; Edwards et al., 1987; Nanson et al., 1991; Auler and Smart, 2001). Ages are determined from the degree of disequilibrium between the higher members of the uranium decay series (Williams et al., 1998, p. 277).



If the sample contains no initial ^{230}Th , any produced during the radioactive decay process will reflect aging during that particular decay pathway. However, if the sample does contain an initial amount of ^{230}Th , certain corrections need to be made when calculating its age. It is also necessary to correct for disequilibrium in the system. In a closed system with no gains or losses of uranium to or from the system, equilibrium occurs when the rate of parent decay and daughter formation are balanced. However, chemical differences between parent and daughter can disrupt this balance, so that the system is in a state of imbalance, or *disequilibrium*. If at any time the material being dated has ceased being a closed system, errors accrue. For example, bone may incorporate (or lose) uranium long after death and burial. Speleothems and tufas may receive detrital thorium as wind-blown dust or in run-off. Corals and mollusc

shells may become diagenetically altered after death, and they may lose or gain uranium from seawater or groundwater, leading to ages that are too young or too old, respectively. Walker (2005, p. 70) considered unaltered coral, clean speleothems and volcanic rocks to be the most reliable materials for U-series dating, with diagenetically altered corals, bone, evaporates, calcretes and peat or wood as generally unreliable. New techniques, whose explanation is beyond the scope of this volume, have enabled greater precision to be obtained. Precision obtained using alpha-particle spectrometry can be as high as 1 per cent and can be improved to under 0.5 per cent using thermal ionisation mass spectrometry (TIMS). Dating speleothems relies increasingly on precise measurements obtained using multi-collector inductively coupled plasma mass spectrometers (MC-ICP-MS), and analytical measuring procedures continue to improve. The useful age range now available from U-series dating is from present up to 500 ka (Table 6.1), which makes this a very versatile dating tool for Quaternary rocks and fossils. As with all dating methods, the wider the array of independent techniques used at any site, the more likely it is that errors will be detected.

TIMS uranium-thorium dating of the $\delta^{13}\text{C}$ record from three stalagmites from caves in southern France and northern Tunisia has revealed synchronous changes with Chinese stalagmite $\delta^{18}\text{O}$ records and important differences with Southern Hemisphere cave records (Genty et al., 2006). The same dating method enabled Denniston et al. (2007) to identify intervals of late Holocene aridity shown in the stalagmite record from a cave in central Missouri. Uranium-series ages obtained on speleothems from South Africa and Somalia have been used to date wet phases in both regions over the past 300 ka and, together with ^{14}C ages spanning the last 35 ka, have shown that when southern Africa was wet in late glacial times, eastern Africa was dry, and vice versa during the early Holocene (Brook et al., 1997). Vaks et al. (2007) used high-precision MC-ICP-MS measurements to obtain U-Th ages for cave deposits (speleothems) from the central and southern Negev Desert of Israel to show that the last interval of sporadically wet climate was from 140 to 110 ka, favouring early modern human dispersal from north-east Africa into the Levant at that time. Another application of the U-series method involved U-Pb dating of water-table indicator speleothems (*cave mammillaries*) from nine sites in the Grand Canyon to reveal that the Grand Canyon had developed by headward erosion from west to east, with accelerated incision in the east at around 3.7 Ma (Polyak et al., 2008).

6.7 Dating methods based on trapped electrons

6.7.1 Luminescence dating of dunes, loess and other quartz-rich sediments

Aitken has provided detailed accounts of the principles and practice of thermoluminescence dating (Aitken, 1985) and of optical dating methods (Aitken, 1998), while Duller (2008) has produced a manual on luminescence dating that is specifically aimed

at archaeologists but is of value to any non-specialist concerned with luminescence dating. Luminescence dating has been in increasing use since the 1960s, when it was widely used to date archaeological remains such as pottery and bricks. The method is based on the fact that certain minerals, such as quartz, feldspar, calcite and zircon, can store energy within their crystal lattice structure as trapped electrons (Aitken, 1985; Aitken, 1990; Lian and Huntley, 2001; Duller, 2004; Duller, 2008). Grains within a sedimentary deposit receive energy from the emissions of minute amounts of radioactive isotopes (mainly ^{40}K , U and Th) in the surrounding material and store it in this way. The grains thus act as natural dosimeters, recording the amount of radioactivity to which they have been exposed (Duller, 2004; Duller, 2008). The application of controlled amounts of heat or light causes the grains to release the stored energy in the form of light, a phenomenon termed *luminescence*. Precise measurement of the brightness of the luminescence signal allows the total amount of radiation to which the sample has been exposed since burial to be calculated. This is done by finding the radiation dose applied in the laboratory that produces the same amount of luminescence as the sample. This is called the *equivalent dose*. Dividing the equivalent dose by the amount of radiation received by the sample each year (*dose rate*) will give the age (Duller, 2008). When heat is applied to release the stored energy, the method is called thermoluminescence (TL) dating, and when light is used, it is called optically stimulated luminescence (OSL) dating.

Huntley et al. (1985) invented OSL dating more than twenty-five years ago, and it is now the most widely used method of luminescence dating (Aitken, 1998), with a possible age range from 10–20 years to more than 500,000 years (Huntley and Prescott, 2001). OSL has been especially successful in dating eolian sediments, such as dunes (Huntley and Prescott, 2001; Singhvi et al., 2010) and loess (Roberts et al., 2003), where the quartz grains have been well-exposed to sunlight before being buried, so the grains have been fully bleached and the luminescence ‘clock’ reset to zero. Results obtained from OSL dating of Nile alluvium are consistent with paired radiocarbon ages from these deposits, inspiring cautious confidence in these dating techniques (Williams et al., 2003; Williams et al., 2010b). However, there is always a possibility that quartz grains laid down in alluvial or lacustrine settings may only be partially bleached. This means that the luminescence energy acquired prior to burial has not been removed, and the measured age will be too large. The use of aliquots (sub-samples) containing only a small number of grains or single grain methods pioneered by Duller (1991; 1995) can enable detection of whether or not the grains have been most fully bleached. Interesting applications of this technique include using single grains to date rates of bioturbation in soils and rates of soil formation (Pillans et al., 1997). There has also been some success in dating feldspars by using infrared radiation to stimulate luminescence, a method termed IRSL dating. For example, the results obtained on dating alluvial samples in the semi-arid Son Valley of north-central India using IRSL appear consistent with the radiocarbon chronology

derived from charcoal and shells from these deposits (Pal et al., 2004; Williams et al., 2006b).

6.7.2 Electron-spin resonance dating of tooth enamel, mollusca and tephra

Electron-spin resonance (ESR) dating is very similar to TL and OSL dating in that it depends on the trapping of electrons, but it is based on direct measurement of the number of trapped electrons rather than the energy they release on stimulation by heat or light (Blackwell, 2001). This method has the advantage of being able to date samples ranging from modern to more than a million years in age that are not amenable to dating using TL or OSL. The precision of ESR dating is of the order of 10–20 per cent but can be as low as 100 per cent, depending on the type of sample. One problem with ESR dating stems from the type of sample being dated, such as tooth enamel or coral, and concerns the possible uptake or loss of uranium in the sample, complicating the calculation of dose rate. However, this can be overcome in part by using different models to estimate dose rate, one assuming a closed system and one assuming an open system for the uranium in the tooth or coral sample under analysis. As with all dating methods, independent verification of the age obtained is important. A great advantage of ESR dating is that it provides a direct age for the fossil being dated (Grün and Stringer, 1991; Grün et al., 2001). Another advantage is that samples are not altered by the application of heat or light, as with luminescence dating, so replicate analyses are possible.

6.8 Dating with cosmogenic radioisotopes

Cosmogenic nuclide dating has really come of age in the last decade or so, and an increasing number of research laboratories around the world are now specialising in this powerful dating technique. When high-energy cosmic rays enter the atmosphere they collide with the nuclei of atmospheric gas atoms, generating a flux of high-energy neutrons and minor amounts of *muons* to the earth's surface. (Muons are unstable elementary particles with a mass 207 times that of an electron, and are negatively charged.) These and other subatomic particles react with certain rock minerals, creating new nuclides. The concentration of these newly created nuclides decreases from the surface down and is a direct function of the time elapsed since the rock surface has been exposed to cosmic radiation, hence the alternative expression *surface exposure dating* for this method of dating. One application of the method is to determine long-term rates of denudation. The assumption here is that over several million years, the mean rate of surface lowering balances the rate of accumulation of cosmogenic nuclides in the near surface, which is a reasonable assumption for tectonically stable uplands in arid areas, as well as for desert pavements on stable land surfaces (Fujioka et al., 2005; Heimsath et al., 2010; Quigley et al., 2010a; Quigley et al.,

2010b; Fujioka and Chappell, 2011). A technically more difficult application is in dating alluvial deposits and beach ridges, such as the 386 m shoreline of the White Nile. Here allowance must be made for bulk density changes and for later burial events, requiring more complex modelling of the primary analytical data (Barrows et al., 2014). Cosmogenic nuclides currently used in exposure dating (with age range in brackets, after Walker, 2005, table 3.2) include the stable nuclides ^3He (1 ka to approximately 3 Ma) and ^{21}Ne (7 ka to approximately 10 Ma) and the unstable nuclides ^{10}Be (3 ka to 4 Ma), ^{26}Al (5 ka to 2 Ma), ^{36}Cl (5 ka to 1 Ma) and, of course, ^{14}C , discussed in some detail earlier in this chapter.

Cosmogenic ^{10}Be has been used with success to date late Quaternary glaciations in the Tian Shan ranges in central Asia (Kong et al., 2009) and to date glacial and periglacial deposits in the semi-arid Snowy Mountains of south-east Australia (Barrows et al., 2001; Barrows et al., 2002; Barrows et al., 2004). Ages of Sierra Nevada cave sediments based on the ratio of aluminium-26 to beryllium-10 ($^{26}\text{Al}/^{10}\text{Be}$) have been used to determine rates of late Pliocene and Quaternary river incision and are consistent with tectonic uplift of the Sierra Nevada during the past 10 million years (Stock et al., 2004). One possible problem with the use of ^{10}Be , however, concerns variations in its production rate. Raisbeck et al. (1985) found that during a geomagnetic reversal, there was an increase in cosmogenic ^{10}Be . During the Last Glacial Maximum, the earth's magnetic field was weaker than it is today, and this was reflected in an increase in the production of cosmogenic ^{10}Be (Lao et al., 1992).

6.9 Chemical methods: amino acid racemisation dating

In one sense, this method may be regarded as a means of obtaining relative ages and so might best be classed as a Category I correlation method. However, technical advances have effected such substantial improvements in both accuracy and precision that it deserves separate treatment. Amino acids are protein molecules with asymmetric carbon bonds and are characterised by alternative molecular patterns, or *isomers* (somewhat akin to mirror images), termed *stereoisomers*. These stereoisomers consist of two distinct types, the left, or L, type and the right, or D, type. Once an organism dies, there is a progressive change from the L type dominant in living molecules to a balanced mixture of L and D types, a process termed *racemisation* for isomers with one asymmetric carbon atom and *epimerisation* for isomers with two such atoms. The rate of change is a linear function of time, but it varies in regard to temperature (Blackwell, 2001). Low temperatures slow down the racemisation rate, while higher temperatures speed it up. Age of the fossil organism, whether shell, wood or bone, is estimated from the relative proportions of L and D isomers, but because the racemisation rate is highly sensitive to temperature, an independent method of assessing temperature is needed if this technique is to provide credible age estimates. Alternatively, the sample in question, for instance, ostrich or emu eggshell, can be dated independently using

AMS ^{14}C or ESR, and the mean temperature prevalent since the organism died can be estimated (Miller et al., 1997). Errors may arise from post-mortem contamination by fungi and bacteria, which add new amino acids to the sample, by amino acids present in groundwater or by diagenetic changes in amino acid composition linked to weathering processes. Provided these potential or actual contaminants are identified, the method can yield usefully accurate results, although they are seldom of high precision. The method has been successfully used to date land snails in the Negev Desert (Goodfriend, 1992), ostrich eggshells associated with late Quaternary archaeological sites in the eastern Sahara (Brooks et al., 1990) and eggshells of the extinct late Pleistocene ratite, *Genyornis*, in central Australia (Kaufman and Miller, 1995).

6.10 Other dating methods

This short review is not a treatise on dating and has perforce passed over a number of other methods of relative and absolute dating of Cenozoic rocks and fossils. One such method is fission track dating, which has been used to date fossil hominids in East Africa (Gleadow, 1980). There have also been a number of problematic and ultimately highly controversial attempts at cation-ratio dating of Quaternary rock varnish and prehistoric rock engravings in arid areas (see Dorn et al., 1989; Dorn, 1990; Dorn et al., 1990; and discussion by Bierman and Gillespie, 1991; Bierman et al., 1991; Reneau and Raymond, 1991; Reneau et al., 1991). Very little is now heard about this technique, but work is quietly underway to use certain trace elements in rock varnish and in silcrete (discussed in Chapter 15) as a dating tool in the future. Other methods that have the potential to provide an annual record involve the now well-established counting of annual growth rings in trees (dendrochronology) (Fritts, 1976) and the counting of annual layers of ice within ice cores, of which the Greenland and Antarctic ice core chronologies are the most widely used (Dansgaard et al., 1984; Dansgaard et al., 1985; Dansgaard et al., 1993).

6.11 Conclusion

The invention of radiocarbon dating by Willard Libby in 1949 revolutionised the study of geologically recent sediments in deserts, allowing river, lake and archaeological deposits to provide credible ages back to about 40,000 years ago. Until then, we had to rely on relative dating methods, such as the correlation of similar sedimentary units or prehistoric stone tool assemblages. Other methods of absolute dating extended the time range well beyond the 40 ka of ^{14}C conventional dating methods. Uranium-series dating allowed speleothems in desert caves to be dated back to 500 ka, and provided less stringent age control for lake and soil carbonates, given the problems of uranium losses from the parent carbonate. Potassium-argon and later argon-argon

dating techniques allowed Miocene and Pliocene fossil hominids to be accurately dated, complementing the use of geomagnetic dating methods, which provided reliable but not very precise ages and, in any event, needed independent calibration. Another revolutionary development was that of luminescence dating, both thermoluminescence and optically stimulated luminescence dating methods. For the first time, it became possible to date directly the quartz particles within desert sand dunes and desert loess deposits to discover when they were last exposed to sunlight. These techniques have provided reliable ages back to approximately 100 ka, under ideal conditions back to approximately 500 ka and, with developments underway, may soon yield ages back to 1 million years ago. Finally, the development of exposure dating methods using a variety of cosmogenic nuclides has provided reliable ages for glacial moraines, desert pavements and fluvial sediments, as well as denudation rates in arid and semi-arid areas. Advances in dating techniques over the past sixty years have thus truly revolutionised the study of climatic change in deserts.

Stable isotope analysis and trace element geochemistry

Se non è vero, è molto ben trovato.

It may not be true, but it is very well contrived.

Giordano Bruno (1548–1600)

De gl'heroici furori (1585)

7.1 Introduction

We saw in [Chapter 6](#) that the different isotopes of a chemical element have the same number of protons but different numbers of neutrons and noted that individual isotopes of an element are termed nuclides. The focus of this chapter is on nuclides that are stable through time, as opposed to those that are unstable and undergo radioactive decay, making them especially valuable for dating rocks and sediments ([Chapter 6](#)). This chapter is not a treatise on stable isotopes – excellent monographs and reviews are available for those seeking more information (Dansgaard, [1964](#); Yurtsever, [1975](#); Pearson and Coplen, [1978](#); Yurtsever and Gat, [1981](#); Faure, [1986](#); Sealy, [1986](#); Hoefs, [1997](#); McDermott, [2004](#); Fairchild et al., [2006](#); Leng and Barker, [2006](#); Leng and Barker, [2007](#); Leng and Sloane, [2008](#)). Rather, it seeks to provide examples of how stable isotopes have been used to reconstruct climatic change in deserts. A great deal of progress has been made in this field during the past forty years. When Flint ([1971](#)) and Butzer ([1971](#)) published their monumental monographs on geologically recent global environmental changes, they hardly mentioned the use of stable isotopes in reconstructing past terrestrial environments. Today, stable isotopes are one of the most widely used indicators of climate change in the ever-growing and impressive arsenal of available tools listed in [Table 1.2](#) of [Chapter 1](#).

The use of stable isotope analysis has proven to be a very powerful tool in reconstructing past environmental fluctuations in deserts, desert margins and the oceans surrounding the deserts. Because carbon, oxygen and hydrogen are abundant in rocks, sediments, plants, rivers and lakes, the stable isotopes of these three elements are the

ones most commonly used to determine former environmental fluctuations in the arid lands and, indeed, elsewhere.

Changes in the stable isotopic composition of carbon, expressed as the ratio of ^{12}C to ^{13}C (or $\delta^{13}\text{C}$) can shed light on past changes in vegetation and in human and animal diet. Changes in the ratio of ^{16}O to ^{18}O (or $\delta^{18}\text{O}$) in organic and inorganic carbonates and in ice have been used to infer past changes in temperature, global ice volume and salinity. Finally, changes in the ratio of light to heavy hydrogen D/H (or δD) in, for example, ice cores, reflect local temperature changes above the ice caps. In fluid inclusions within speleothems, the D/H ratio reflects the ambient temperature within the cave.

Another potentially significant advance is the use of ‘clumped isotopes’ to determine past temperature changes, although this technique is still very much in its infancy and will need considerable independent testing and calibration (Ghosh et al., 2006a). Clumped isotopes are those rare isotopes in which the concentration of ^{13}C - ^{18}O bonds in reactant carbonate is a function of temperature at the time of carbonate growth. The method has also been used to determine the uplift rates of the Altiplano plateau of the Bolivian Andes by analysing the concentration of ^{13}C - ^{18}O bonds in paleosol carbonates (Ghosh et al., 2006b). Inferred rates amounted to 1.03 ± 0.12 mm/year between approximately 10.3 and approximately 6.7 Ma. Uplift rates over the past two decades measured from space geodetic observations amount to around 10 mm/year, indicating an order of magnitude increase in uplift rate since the late Miocene (Fialko and Pearce, 2012). Clumped isotopes have also been used to determine speleothem temperature in Soreq Cave (Affek et al., 2008) and can be used to determine the temperature to within $\pm 2^\circ\text{C}$ in any carbonate precipitated in equilibrium with its surrounding environment. This would preclude speleothems in caves where evaporation rates were high and variable.

Another set of isotopes that have seen increasing use are strontium isotopes; the strontium isotope ratio $^{87}\text{Sr}/^{86}\text{Sr}$ preserved in alluvial clays and aquatic snail shells is used to reconstruct past changes in river discharge, as well as to fingerprint individual layers of volcanic ash. In combination with the neodymium isotope ratios $^{143}\text{Nd}/^{144}\text{Nd}$, the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios found in wind-blown dust, as well as in reworked desert loess and associated soil carbonate nodules, have been used to identify dust source areas. Similarly, the isotopic ratio of carbon to nitrogen (C/N) provides an indication of the diet of prehistoric (and modern) mammals, including humans. Examples of each of these applications are given in the subsequent sections. In addition to the use of stable isotopes, there have been significant advances in the use of trace element geochemistry to determine past changes in temperature and salinity in lakes, as well as likely sources of desert dust. Trace element geochemistry is also routinely used to fingerprint volcanic ash beds and thus to determine the parent volcanic source of the ash (Shane et al., 1995; Shane et al., 1996; Westgate et al., 1998).

7.2 Factors that influence the stable isotopic composition of water, plants and other organisms

The isotopic composition of desert rainfall will depend on a number of independent factors, including latitude, elevation, total distance travelled from the initial source of moist air, evaporation in transit and rainfall losses en route to the final destination (Dansgaard, 1964). Other things being equal, the greater the evaporation in transit and the greater the losses of water vapour before the final precipitation event, the higher the proportion of ^{18}O in the rain, and this will be reflected in the $\delta^{18}\text{O}$ values measured in rainfall samples. Evaporation is hard to assess accurately, and none of the existing models are fool-proof, including the classic Rayleigh distillation models that have been in use for more than a century (Rayleigh, 1896; Hoefs, 1997). Further changes in stable isotopic composition occur as a result of isotopic fractionation linked to both biological and kinetic effects once the rain has percolated through the soil and reached the groundwater table. Groundwater flowing into lakes, springs and rivers will also reflect interaction with the parent rock through which it flows (Wigley, 1976), and this may influence the $\delta^{13}\text{C}$ values of the water. It is important to bear in mind that many non-climatic factors will affect isotopic composition, including the influence of bedrock on both $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$. Variations in the types of plants growing on the surface and of the various photosynthetic pathways they follow will have a direct influence on $\delta^{13}\text{C}$, as will plant respiration and bacterial activity within the soil. These particular influences are discussed in [Section 7.7](#).

7.3 The marine oxygen isotope record

This topic was covered in [Chapter 6](#), so only a few brief comments are needed here. When water evaporates from the surface of the ocean, the lighter isotope of oxygen (^{16}O) is preferentially removed and the surface waters of the ocean become enriched in the heavier isotope (^{18}O). With evaporation of water vapour from the ocean surface, the sea surface becomes more saline. The $^{16}\text{O}/^{18}\text{O}$ ratio in the calcareous shells and tests of marine microorganisms (especially marine foraminifera) are thus measures of near-surface ocean salinity, as recognised by Emiliani (1955) more than half a century ago. Duplessy (1982) interpreted the increase in ocean surface salinity in the northern Bay of Bengal during the last glacial as a reflection of reduced freshwater discharge into the Bay of Bengal as a result of enhanced glacial aridity, a conclusion consistent with the presence of 25–15 ka carbonate cemented alluvium in the Son Valley of north-central India (Williams and Clarke, 1984; Williams and Clarke, 1995). The $^{16}\text{O}/^{18}\text{O}$ ratio in ocean water and in the calcareous shells of marine organisms is also a measure of ocean water temperature, with high rates of evaporation under warmer conditions and low rates under cooler ones, although other factors such as wind speed will also influence evaporation from the ocean surface. At a local and even regional

level, changes in the $^{16}\text{O}/^{18}\text{O}$ ratio within calcareous marine fossils can be used as a measure of past changes in near-surface sea salinity and temperature.

However, the overwhelming factor controlling fluctuations in this ratio is the waxing and waning of the great continental ice sheets, as Shackleton (1967; 1977; 1987) was the first to recognise. As progressively more water is abstracted from the ocean and incorporated into the ice caps, the ocean becomes more enriched in the heavier isotope, and this is reflected in the $^{16}\text{O}/^{18}\text{O}$ ratio. The effect is so pronounced across the oceans of the Earth that fluctuations in the ratio of ^{16}O to ^{18}O , expressed as $\delta^{18}\text{O}$ in parts per thousand (‰, or ‘per mil’), have been used to reconstruct a detailed relative chronology (see Chapter 6 for details) of what are now the well-known Marine Isotope Stages, with the interglacial phases numbered back from the Holocene interglacial (MIS 1) using uneven numbers and the glacial phases numbered back from MIS 2 (incorporating the Last Glacial Maximum at 21 ka) using even numbers. A pair of marine isotope stages will, in principle, coincide with a single, complete glacial-interglacial cycle, but complications arise from the inclusion of interstadial and stadial episodes within the MIS chronology, so they should be discussed as separate entities. Imbrie et al. (1984) provided a revised version of the marine chronology, known as the SPECMAP $\delta^{18}\text{O}$ record, and it has been widely accepted since then.

A further issue is the need to calibrate the MIS relative chronology against a measure of absolute time, for which the preferred templates are the various orbital cycles (Chapter 6), a process termed ‘orbital tuning’. Here again, difficulties arise over possible time lags between inferred insolation changes at various latitudes linked to the changes in orbital geometry, such as the distance of the earth from the sun or the tilt of the earth’s axis, and the response of the continental ice sheets, which may involve time lags of thousands of years. Ideally, any use of the marine isotope stage chronology should be supplemented by independent dating of the relevant marine core using the methods described in Chapter 6. In the absence of any such age control, the use of ‘wiggle matching’ between sets of MIS curves (e.g., Lisiecki and Raymo, 2005; Lisiecki and Raymo, 2007; Raymo and Huybers, 2008) and other evidence of environmental change should be carried out circumspectly.

7.4 The oxygen and carbon isotope record in desert lakes and rivers

Analysis of the stable isotope ratios of oxygen and carbon in carbonates from desert lakes, swamps and playas has shed useful light on past changes in water depth, temperature and chemistry (Fontes and Pouchan, 1975; Cerling et al., 1977; Cerling, 1979; Fontes et al., 1983; Lemeille et al., 1983; Gasse and Fontes, 1989). In the case of coastal sebkhas, or salt pans, it has also been used to determine past fluctuations in sea level (Fontes and Perthuisot, 1971).

Subsequent isotopic work has focussed rather more on the freshwater gastropod and ostracod shells within lake and river sediments rather than the sedimentary

carbonates themselves, which often only give a very generalised climatic signal (Abell, 1985; Williams et al., 1987; Abell and Williams, 1989; Eyles and Schwarcz, 1991; Abell et al., 1996; Ayliffe et al., 1996; Abell and Hoelzmann, 2000; Glasby et al., 2007).

As a precursor to analysing the isotopic ratios in freshwater gastropod shells from the Sahara, Nile Valley and East Africa, Abell (1985) compiled the isotopic ratios in modern gastropod shells from more than eighty localities throughout Africa. The analysed values showed consistent variations with latitude, elevation and the amount of rainfall, in good accord with previous compilations of the isotopic composition of rainfall across the globe (Dansgaard, 1964; Yurtsever, 1975). Abell and Williams (1989) examined the oxygen and carbon isotope ratios in gastropod shells recovered from Holocene lake and spring sediments in the southern Afar and Ethiopian rifts. They also examined isotopic changes along the growth whorls of *Melanoides tuberculata* shells living in Lake Lyadu, a small lake in the southern Afar, which dried out soon after the shells had been collected. The fluctuations revealed intervals of little or no growth, probably as a result of a sharp drop in lake level during the recent droughts in that area. Results from the other Ethiopian samples were consistent with the known history of lake level fluctuations in that region, and they revealed a clear discrimination between gastropods that had lived in ponds fed primarily from local springs and lakes in receipt of far-travelled run-off from the Ethiopian Highlands.

Fontes et al. (1985) investigated the diatom assemblages and stable isotopic composition of Holocene lake sediments in the northern Sahara and found very rapid changes in water chemistry, from fresh to highly saline, within this time interval. At the site of Adrar Bous in the Ténéré Desert of the south-central Sahara, an early Holocene lake associated with Mesolithic remains dried up and was succeeded by a shallower lake associated with Neolithic artefacts, food remains, human burials and the complete skeleton of a domesticated short-horned cow (*Bos brachyceros*). Gastropod shells from the Neolithic lake and charcoal from the Neolithic sites had radiocarbon ages between approximately 6 ka and approximately 4.5 ka (Williams et al., 1987; Williams, 2008). Analysis of the stable carbon and oxygen isotopic composition of the shells indicated some degree of seasonal variability and possibly cooler temperatures but certainly wetter and less evaporative conditions at that time. Another important result was the differential response shown by several different species of gastropod, indicating the need for caution in this type of research.

Ayliffe et al. (1996) analysed the stable carbon and oxygen isotopic composition of gastropods that lived in shallow ponds in the present desert west of the lower White Nile during the early Holocene (Williams et al., 1974). They concluded from the highly negative $\delta^{18}\text{O}$ values of the shells that the region experienced less evaporation at that time and inferred from the extreme variability in isotopic composition (up to 6–7‰ PDB) that there had been considerable interannual variability in precipitation.

(Chapter 6 defines PDB.) Ten new AMS ^{14}C ages obtained on shells from two of the clay pans have confirmed that the interval between 9.9 ka and 7.6 ka, and especially the 600-year interval from 9.0 ka to 8.4 ka, was perhaps three times wetter than today (Williams and Jacobsen, 2011). These ages are comparable to the ages of the Mesolithic barbed bone harpoon sites of Tagra and Shabona east of the lower White Nile (Adamson et al., 1974; Adamson et al., 1987a), as well as the age of a recently mapped 450 km² lake that was fed by an overflow channel from the main Nile in presently arid northern Sudan between 9.5 ka and 7.5 ka (Williams et al., 2010b). Both Blue and White Nile floods were also high during this time (Williams, 2009b), which suggests that regionally wetter conditions in the eastern Sahara away from the Nile coincided with times of high Nile flow in the early Holocene (see Chapter 10).

Lézine et al. (1990) investigated the stable carbon and oxygen isotope stratigraphy and associated pollen grains in a sediment core from Chemchane sebkha in Mauritania that had originally been studied by Chamard (1973). The present salt pan, or sebkha, was a relatively deep freshwater lake between approximately 8.3 and 6.5 ka, at which time there was a 400–500 km northward shift of savanna vegetation. This shift of the early Holocene savanna vegetation belt is consistent with that inferred by Ritchie et al. (1985) and Ritchie and Haynes (1987) in northern Sudan some 3,000 km further east, suggesting that the vegetation may have migrated northwards along the entire southern Sahara.

7.5 The oxygen, carbon and hydrogen isotope record in desert speleothems and tufas

The results of $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ and δD (D/H) analyses of speleothems and, in some instances, tufa deposits from arid, semi-arid and seasonally wet localities in the Americas, China, peninsular Arabia, southern Africa and Israel are covered in detail in Chapter 14. As a result, it is only necessary to consider very briefly some of the assumptions underlying the interpretations of past environments and climatic changes presented in that chapter. Because the $\delta^{18}\text{O}$ values in cave speleothems do not afford an unequivocal temperature signal and can vary with the source of rain water (Vaks et al., 2006) and with rain shadow effects (Vaks et al., 2003), it is always useful to provide an independent quantitative assessment of mean annual cave temperature by analysing, for example, the δD fluctuations recorded in fluid inclusions within the speleothems being studied (Matthews et al., 2000; McGarry et al., 2004). Another promising (but very time-consuming) approach may be to use ‘clumped isotope’ thermometry (Affek et al., 2008). In many cases, none of these methods has been used for independent validation of temperature estimates and until they have, temperatures inferred from the $\delta^{18}\text{O}$ fluctuations will remain qualitative inferences.

The $\delta^{13}\text{C}$ values in cave speleothems are an indirect guide to the type of vegetation growing above the cave, which will also influence the $\delta^{13}\text{C}$ in soil carbonates. Given that the type of vegetation growing in arid areas reflects the amount of precipitation as well as the soil type, it can be used as an indirect measure of effective precipitation. In a novel approach to the question of megafaunal extinctions in North America, Polyak et al. (2012) used the $\delta^{13}\text{C}$ and $\delta^{234}\text{U}$ values in speleothem calcite from Fort Stanton Cave in southern New Mexico as a proxy for effective precipitation. Ideally, some independent control involving pollen analysis, tree ring data or vegetation remains found in packrat or stick-nest rat middens should be used to evaluate inferences based solely on speleothem isotopic data. Such an approach can also yield additional paleoclimatic insights. For example, Marino et al. (1992) used the $\delta^{13}\text{C}$ changes in the C_4 shrub *Atriplex confertifolia* from packrat middens in the western United States to provide a record of glacial to interglacial changes in atmospheric carbon dioxide (CO_2). They found that the atmospheric CO_2 was isotopically lighter during the last glacial period relative to interglacial periods, probably as a result of a reduced terrestrial biomass and lower biological productivity in the polar oceans.

7.6 The strontium isotope record in desert lakes, rivers, dust and volcanic ash

On occasion, the results from one line of analysis do not accord with what appears to have been well-established. For example, radiocarbon dates obtained from freshwater gastropod shells in alluvial sediments along the lower White Nile indicated that the most recently abandoned flood-plain of that river began to form as far back as 14–15 ka (Adamson et al., 1980; Williams and Adamson, 1980; Adamson et al., 1982; Williams and Adamson, 1982). Pollen and glaciological studies from the Ugandan highlands also indicated a postglacial rise in temperature and precipitation after 14–15 ka, consistent with diatom and sedimentary evidence of overflow from Lakes Victoria and Albert into the upper White Nile in southern Sudan (Livingstone, 1980). In a similar vein, the Blue Nile began to deposit clay across its flood-plain soon after 15 ka, after a long dry interval during the Last Glacial Maximum, when lakes were dry throughout East Africa (Butzer, 1980; Adamson et al., 1980; Williams and Adamson, 1982; Adamson et al., 1982). This substantial body of work was ignored in a short paper claiming that Lake Victoria had remained a closed basin until 7.2 ka (Beuning et al., 1997a). This conclusion was inferred from $^{18}\text{O}/^{16}\text{O}$ ratios in sediment cellulose from Lake Victoria. It was contrary to three decades of research throughout the Nile Basin and ran counter to what was already well-established in Nile Valley prehistoric archaeology (Butzer and Hansen, 1968; Butzer, 1980).

Talbot et al. (2000) resolved to test this hypothesis and to determine more precisely when Lake Victoria began to overflow after its long dry interval during the Last Glacial Maximum. They used strontium isotopes as tracers to ascertain just when Lakes

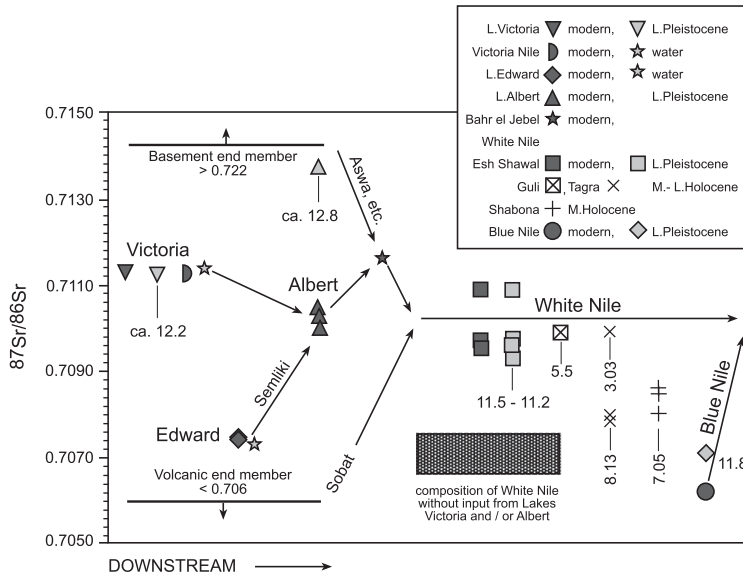


Figure 7.1. Strontium isotopic composition of Blue and White Nile waters and of lakes in the White Nile headwaters. The numbers are radiocarbon ages of dated late Quaternary samples from sites along the Blue and White Nile valleys. (After Talbot et al., 2000.)

Victoria and Albert overflowed into the upper White Nile. They also analysed the strontium isotope ratio ($^{87}\text{Sr}/^{86}\text{Sr}$) preserved in freshwater gastropod shells collected from Blue and White Nile sediments ranging in age from terminal Pleistocene to present-day and compared the values obtained with the strontium isotope ratios from the Ugandan lakes (Figure 7.1). Because these ratios are not changed by weathering and fluctuating hydrological cycles, the strontium ratios of river and lake waters give a weighted average for the type of rocks within the various basins making up the overall river system. For example, in volcanic catchments, the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios will generally be <0.706 , and in those on Precambrian metamorphic basement complex rocks, the ratios will be >0.722 (Williams et al., 2006c). All of the shells analysed had been tested for carbonate recrystallization using X-ray diffraction. The results showed that overflow from Lake Victoria was underway by around 14.5 ka and provided an independent confirmation that the present-day integrated Nile drainage network became re-established at that time, a conclusion confirmed by later studies (Williams et al., 2000; Johnson et al., 2000; Williams et al., 2006c; Williams 2009b; Williams et al., 2010b).

Fluctuations in the $^{87}\text{Sr}/^{86}\text{Sr}$ values preserved in Nile Delta clays have been used to reconstruct a Holocene history for the main Nile (Krom et al., 2002; Stanley et al., 2003; Woodward et al., 2007), as well as for Lake Albert (Williams et al., 2006c). One dramatic event recorded by the strontium isotopic data is the sudden collapse of

the Old Kingdom in Egypt brought about by the catastrophic drought centred around 4.2 ka and evident as far away as Mesopotamia and the Indus Valley. In strong contrast to the Egyptian Old Kingdom, the Kerma civilization in the Nile Valley of northern Sudan survived this event (evident in the strontium ratios in local Nile alluvium) and persisted for a further thousand years before succumbing to invasion from Egypt and a further decline in Nile flow (Macklin et al., 2013).

Other uses of strontium isotopes include determining the sources of alluvial clays, calcareous dust (Dart et al., 2004; Dart et al., 2007), volcanic ash, wood, shells and bones, including fish vertebrae. In the case of wind-blown dust, strontium isotope analysis is often combined with analysis of the neodymium isotope ratios $^{143}\text{Nd}/^{144}\text{Nd}$ to pinpoint dust provenance more exactly (Dart et al., 2007). Chen et al. (1999) compared the Rb/Sr ratios measured in two loess profiles in central China covering the last 130 ka with the marine SPECMAP $\delta^{18}\text{O}$ curve of Imbrie et al. (1984), which is a measure of changes in global ice volume, as explained in Section 7.3. They found a close correlation between the two and concluded that the Rb/Sr ratio is a sensitive indicator of changes in the East Asian monsoon associated with changes in global ice volume.

Initial determination that the late Quaternary volcanic ash found in the Son Valley of north-central India by Williams and Royce (1982) had come from the most recent (74 ka) eruption of Toba volcano in Sumatra arose from a comparison of the $^{87}\text{Sr}/^{86}\text{Sr}$ values preserved within the ash (Williams and Clarke, 1995) with those obtained by Whitford (1975) from welded tuffs around the parent caldera.

7.7 The carbon isotope record in fossil plants, soils, bones and teeth

During photosynthesis, plants absorb carbon dioxide from the atmosphere and, under the influence of sunlight and the plant enzyme chlorophyll, convert the CO_2 to starch, which is then used for plant growth. Plants fix atmospheric carbon dioxide in one of three different ways. All trees, most shrubs and those grasses that grow in shaded forests or in temperate regions follow the Calvin, or C_3 , pathway of photosynthesis (van der Merwe, 1982; Cerling et al., 1991). Grasses adapted to growing in strong sunlight, such as most tropical grasses, follow the Hatch-Slack, or C_4 , photosynthetic pathway. Most succulent plants follow the third, or CAM, pathway, so named because it involves crassulacean acid metabolism, abbreviated to CAM. All three systems of photosynthesis fractionate the carbon isotope ratio of atmospheric carbon dioxide in quite different ways (Vogel, 1978; Vogel et al., 1978; van der Merwe, 1982).

As a result of this differential carbon isotopic fractionation during photosynthesis, C_4 plants tend to have $\delta^{13}\text{C}$ values between -9‰ and -16‰ (mean: -12.5‰), C_3 plants have $\delta^{13}\text{C}$ values between -20‰ and -35 per mil‰ (mean: -16.5‰) and CAM plants have $\delta^{13}\text{C}$ values averaging about -16.5‰ (van der Merwe, 1982). In controlled environment growth cabinets, Read and Farquhar (1991) examined the

differences in carbon isotope discrimination in the leaves of twenty-two species of *Nothofagus* trees from across the Southern Hemisphere. They found that such long-term and genetically controlled discrimination provided a measure of water use efficiency by the leaves, and therefore of their comparative ability to adapt to water stress.

One complication can arise as a result of biogenic recycling of atmospheric CO_2 in certain environments such as flood-plains, leading to carbon isotope gradients varying across the same ecosystem, both between different species and within the same species (Martinelli et al., 1991).

Further fractionation occurs when animals eat the plants. Bone collagen becomes enriched by about 5‰ relative to the average $\delta^{13}\text{C}$ value of the plants ingested. This can then indicate what type of plants the animals were eating. In South Africa, where such studies were pioneered in the late 1970s, browsing animals such as kudu only feed on leaves of trees and shrubs, which are C_3 plants. The average $\delta^{13}\text{C}$ value of the bone collagen of such animals is around -21.5‰ (van der Merwe, 1982). Grazing animals, which eat only C_4 grasses, have $\delta^{13}\text{C}$ values between -8‰ and -10‰ , and mixed feeders, such as sable antelopes, have values of -13‰ to -15‰ .

Using these empirically determined quantitative data, it is possible to use isotopic measurements on bones and teeth to assess the relative amounts of C_3 and C_4 plants eaten by prehistoric animals and people (Sealy, 1986). A refinement of this approach is to use both carbon and nitrogen bone collagen ratios to determine prehistoric human diets, a method that can discriminate between cereals, fish and even between camel herders and goat, sheep and cattle herders (Ambrose and DeNiro, 1986), although climate also has an influence on the isotopic composition of bone nitrogen and must therefore be taken into account (Heaton et al., 1986). Using carbon isotopes preserved in the Miocene/Pliocene fossil fauna of the Potwar Plateau of Pakistan and the Tugen Hills in northern Kenya, Morgan et al. (1994) were able to show that C_4 grasses had first appeared in the animal diet by 9.4 Ma and were present in Kenya by 15.3 Ma, although not as a major component of herbivore diet until 7 Ma.

One practical by-product of this work has been the ability to use a combination of isotope ratios ($^{13}\text{C}/^{12}\text{C}$, $^{15}\text{N}/^{14}\text{N}$ and $^{87}\text{Sr}/^{86}\text{Sr}$) to determine the source areas of elephant ivory, thereby providing a powerful means to help control the illegal trading in ivory (van der Merwe et al., 1990; Vogel et al., 1990).

Ayliffe and Chivas (1990) investigated the oxygen isotopic composition ($\delta^{18}\text{O}$) of bone phosphate ($\delta^{18}\text{O}_\text{p}$) in modern kangaroos and wallabies (Macropods) from a variety of climatic zones across Australia. They found a strong correlation between $\delta^{18}\text{O}_\text{p}$ and mean annual environmental relative humidity. Since Macropods consume substantial amounts of plant leaf water, the $\delta^{18}\text{O}_\text{p}$ probably reflects leaf water fractionation processes, which are in turn controlled by relative humidity. They concluded that $\delta^{18}\text{O}_\text{p}$ from fossil Macropod bones could be used as a measure of past changes in humidity. On the face of it, this seems to be a reasonable conclusion. However, the stability of apatite phosphate in fossil bones and teeth, long considered resistant

to change, may in fact show microbially induced alteration soon after death (Zazzo et al., 2004). This caveat would also apply to claims that $\delta^{18}\text{O}$ variations measured on samples of dentinal hydroxyapatite collected from annual growth bands in the tusks and teeth of late Pleistocene mastodons and mammoths can be used to measure seasons of death and seasonal paleoclimates (Koch et al., 1989).

Cerling et al. (1991) used the stable carbon isotopic composition of pedogenic carbonate and organic matter from paleosols at a 14 Ma site in western Kenya with abundant remains of fossil fauna, including hominoids, to test competing hypotheses about the plant cover at that time. (The Hominoidea superfamily includes the apes and ancestral humans, as discussed in Chapter 17). They were able to demonstrate conclusively that C_3 plants dominated the local vegetation when the soils were forming, so that the soils had probably developed under forest or woodland. Any of the grasses in these soils identified by earlier workers were most likely transient features of the landscape that had developed in the aftermath of volcanic eruptions but did not persist for long and so had little impact on soil development.

Ambrose and Sikes (1991) studied the $^{13}\text{C}/^{12}\text{C}$ ratios in soil organic matter in late Holocene soils along an altitudinal transect in the central Rift Valley of Kenya. They found that the forest-savanna boundary had advanced by more than 300 m in altitude, an observation not evident from earlier studies of the regional pollen record.

Analysis of stable carbon isotopes in paleosols associated with 4.4 Ma *Ardipithecus ramidus* hominid fossils at Aramis in the Afar Desert of Ethiopia has been used to deduce in what type of environment these early hominids might have lived (WoldeGabriel et al., 2009). Additional isotopic investigations included analysis of oxygen and carbon isotopes in mammalian tooth enamel, supplemented by pollen analysis and phytolith abundance measurements (see Chapters 16 and 17 for more details). Although WoldeGabriel et al. (2009) concluded that the habitat at Aramis consisted of woodland and forest patches, a reappraisal of their primary data by Cerling et al. (2010) reached a different conclusion – namely, that the vegetation comprised tree- or bush-savanna, with 25 percent or less of woody canopy cover. White et al. (2010) in turn offered a vigorous rebuttal of this conclusion and defended the original reconstruction.

Seventy-four thousand years ago (74 ka), the entire Indian subcontinent was covered in a thin layer of volcanic ash 10–15 cm thick derived from the highly explosive eruption of Toba volcano in Sumatra, Indonesia. A key and still hotly disputed question relating to this eruption concerns its possible impact on regional vegetation and climate. In order to help resolve this issue, the $\delta^{13}\text{C}$ values in pedogenic carbonate nodules in paleosols above and beneath the 74 ka ash layer were analysed across a 400 km transect in north-central India (Ambrose et al., 2007a; Williams et al., 2009a). The results showed that before the eruption, this part of India was under woodland or forest and was replaced by open woodland and grassland after the eruption. These results were consistent with pollen analyses carried out on a marine core from the Bay

of Bengal in the northern Indian Ocean, sampled at close intervals above and beneath the 74 ka ash layer in the core (Williams et al., 2009a).

Two other instances of stable isotope analysis of pedogenic carbonate in semi-arid India come from the Thar Desert of Rajasthan in north-west India (Andrews et al., 1998; Singhvi et al., 2010). Andrews et al. (1998) studied pedogenic carbonates within a 70 ka sequence of eolian sands and found that $\delta^{13}\text{C}$ values were highest when the $\delta^{18}\text{O}$ values indicated the most arid conditions. They concluded that during glacial periods when the atmospheric carbon dioxide concentration ($p\text{CO}_2$) was lowered, C_4 grasses expanded at the expense of C_3 plants. Singhvi et al. (2010) found that carbon isotopes measured on organic matter within the sand profiles of a 200 ka polygenic dune profile showed consistent values close to $-21.6 \pm 1\%$, pointing to deposition of the eolian sands during a transitional climatic regime characterised by a change from open C_3 grassland to C_4 woodland or forest. The assumption that sand deposition had occurred during glacial maxima, implicit in earlier studies, was thus replaced with a more nuanced interpretation of maximum dune activity during the transition from weak to stronger summer monsoon in immediate postglacial time. In this case, wind strength outweighed that of aridity in promoting sand movement, as discussed in Chapter 8.

An interesting question is whether or not changes in vegetation caused by changes in $p\text{CO}_2$ can be discerned from analysis of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values in pedogenic carbonates from arid areas. Cole and Monger (1994) analysed paleosol carbon isotope ratios on an alluvial fan in the Chihuahuan Desert of New Mexico and found a shift from mainly C_4 grasses to mainly C_3 shrubs from 9 to 7 ka. This shift coincided with an increase in $p\text{CO}_2$ in Antarctic ice cores and increased local aridity evident from packrat middens. Because the $\delta^{18}\text{O}$ values, which depend on both moisture and temperature, were constant from 9 to 7 ka, when the plant cover changed, they concluded that atmospheric CO_2 rather than regional climate change was the main cause. From this, they also concluded that carbon isotope ratios in ancient soils could be used as a proxy for past changes in atmospheric CO_2 .

A variant on the theme of using carbon isotope ratios in pedogenic carbonate to reconstruct past changes in desert ecosystems involves using land snail shells. Analysis of the $^{13}\text{C}/^{12}\text{C}$ ratios in the organic matrix of fossil land snail shells from the Negev Desert was used to map the past distribution of C_4 shrubs in this region (Goodfriend, 1988). The C_4 shrubs are mainly restricted to the arid zone with less than 280 mm of mean annual rainfall, so shifts in rainfall distribution can be inferred from shifts in vegetation, as inferred from the snail shells. The results indicated that the northern limit of C_4 shrubs had moved south by 20–30 km from approximately 4.5 ka to 2.9 ka, indicating that the northern Negev was far wetter at that time. A further refinement in snail shell analysis involves using the $\delta^{18}\text{O}$ values in dated snail shell carbonate to infer past changes in rainfall and rainfall source area. Goodfriend (1991) obtained seventy-six radiocarbon ages from the Holocene fossil land snail *Trochoidea*

seetzeni from the northern Negev and found that early Holocene $\delta^{18}\text{O}$ values were similar to those of today but were depleted relative to present between 7.4 ka and 6.9 ka, suggesting a change in atmospheric circulation at that time. He attributed this change to an increase in the frequency of storms reaching the Negev from north-east Africa. Modern conditions had resumed by 3.8 ka. (The quoted ages are calibrated ages; the actual published ages were expressed as ^{14}C years BP).

7.8 Trace element geochemistry of ostracods and aquatic snail shells

Initial efforts to unravel the climatic history of desert lakes and wetlands used a variety of methods, including lake sediments, invertebrate fossils such as mollusc and ostracod shells, and any associated pollen grains. Ostracods are very small crustaceans with bivalve shells made of low Mg-calcite. They shed their shells during growth, and the shells are readily preserved in lake muds. In the early 1980s, a new approach was pioneered which involved using the trace element geochemistry as well as the stable isotope analysis of fossil ostracod shells recovered from lake and river sediments (Chivas et al., 1986a; Chivas et al., 1986b; Chivas et al., 1986c). In essence, the Sr/Ca ratio in ostracod shells provides a measure of lake water salinity, and the Mg/Ca ratio provides a measure of lake water temperature. Care needs to be taken to use ostracods belonging to a single species, because different species may display different responses to lake water chemistry (Ito et al., 2003). If the Sr/Ca values cluster tightly within a given sediment core, it shows that salinity did not fluctuate much during the life of the ostracods. If both the Sr/Ca and Mg/Ca values cluster tightly, then lake salinity and temperature did not fluctuate, indicating deep-water conditions (Ito et al., 2003). If both the Sr/Ca and Mg/Ca values fluctuate widely, salinity fluctuations are likely, and the water was either shallow or seasonally fluctuating, as in the last interglacial lakes in the Western Desert of Egypt associated with Middle Stone Age sites (De Deckker and Williams, 1993).

Ito and Forester (2009) expressed reservations about using the Sr/Ca and Mg/Ca ratios in ostracod shells as a measure of water temperature and salinity and suggest that they should be used instead to indicate when changes occur and not why they occur. Many other factors besides salinity will affect algae and other organisms, such as ostracods, that live in lakes, so salinity values inferred from Mg/Ca ratios are best considered as very general estimates to be used in conjunction with estimates derived from species assemblages of diatoms, ostracods and gastropods, as reviewed in Chapter 16.

Salinity may have a negative influence on the size of aquatic organisms. *Melanoides tuberculata* shells observed by the author living today in very shallow saline lakes in the Afar, such as Lake Abhe, were all very small compared to their early Holocene and Pleistocene counterparts, which flourished when this lake was tens of metres deep and fresh.

7.9 Conclusion

Painstaking analysis of the fluctuations in the stable oxygen isotopic composition in many hundreds of marine sediment cores has revolutionised our understanding of past climatic fluctuations, notably in global ice volume, ocean temperature and sea surface salinity. On land, the greatest breakthroughs have come from analyses of $^{18}\text{O}/^{16}\text{O}$ and $^{13}\text{C}/^{12}\text{C}$ ratios in speleothems from caves in Eurasia, Africa and the Americas, as well as from desert lakes across the world. The stable carbon isotopic composition of calcium carbonate concretions in paleosols provides a record of past changes in vegetation, notably the proportions of plants following C_3 and C_4 photosynthetic pathways. Analysis of the C/N ratios in tooth enamel can be used to determine prehistoric human diets, and a combination of $^{13}\text{C}/^{12}\text{C}$, $^{15}\text{N}/^{14}\text{N}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ analyses can be used to determine the provenance of, for example, elephant ivory used in the illegal international ivory trade. $^{87}\text{Sr}/^{86}\text{Sr}$ analysis of freshwater gastropod shells and of fluvial and lacustrine clays is a powerful means of reconstructing past hydrologic and depositional fluctuations and extreme paleoclimatic events, such as the extreme 4.2 ka drought in the north-east quadrant of Africa and beyond. There have been some useful pioneering attempts to use the Sr/Ca and Mg/Ca ratios preserved in fossil ostracod shells as a measure of former lake water salinity and temperature, respectively, although recent studies have stressed the need for greater caution in using these ratios. These studies point out that there are many factors operating within lake waters that influence how organisms respond to possible changes in salinity and temperature.

8

Desert dunes

At times, especially on a still evening after a windy day, the dunes emit, suddenly, spontaneously, and for many minutes, a low-pitched sound so penetrating that normal speech can be heard only with difficulty.

R.A. Bagnold (1896–1990)

The Physics of Blown Sand and Desert Dunes (1941, p. xxi)

8.1 Introduction

Dunes are widely regarded as the quintessentially diagnostic desert landform (Figure 8.1) and have attracted rather more attention than other far more extensive and often potentially more informative desert landforms, such as mountains, pediments, gravel plains, lake basins, alluvial fans and river sediments (Bagnold, 1941; Monod, 1958; Mabbutt, 1968; Cooke and Warren, 1973; McKee, 1979; Wasson, 1984; Lancaster, 1989; Thomas, 1989; Pye and Tsoar, 1990; Yang, 1991; Cooke et al., 1993; Pye and Lancaster, 1993; Abrahams and Parsons, 1994; Lancaster, 1995; Thomas, 1997; Alsharan et al., 1998; Goudie et al., 1999; Parsons and Abrahams, 2009; Yang et al., 2011a; Yang et al., 2011b; Warren, 2013). One reason for this emphasis stems from the extraordinary variations in dune morphology, dune height and dune length, ranging from linear dunes that extend unbroken for hundreds of kilometres to small fields of crescentic barchan dunes advancing downwind to complex individual star dunes that in the Namib Desert are up to 300 m high and in the Badain Jaran Desert of Inner Mongolia attain relative elevations approaching 460 m (Yang et al., 2011a). A second reason is that certain dune forms very closely reflect the direction of the dominant sand-transporting winds. For example, linear dunes appear to run more or less parallel to the dominant sand-moving winds, while the horns of barchan dunes are elongated downwind and parallel to the orientation of the main sand-transporting winds. Given an adequate sand supply and suitable winds, barchans may develop into linear dunes (Figure 8.2). A third reason why dunes have attracted such attention is



Figure 8.1. Sand dunes immediately east of the Aïr Mountains, south-central Sahara. (Photo: J. D. Clark.)

the sheer beauty and symmetry of mobile dunes – a feature commented on at length by the doyen of dune studies, Ralph Bagnold, during his travels in the Libyan Desert in the 1920s and early 1930s (Bagnold, 1935), in his unsurpassed account of *The Physics of Blown Sand and Desert Dunes* (Bagnold, 1941) and in his very readable autobiography (Bagnold, 1990). The focus of this chapter is on the utility of dunes as paleoclimatic indicators. For details of the dynamics of eolian sand movement and more comprehensive accounts of desert dunes, the interested reader should consult

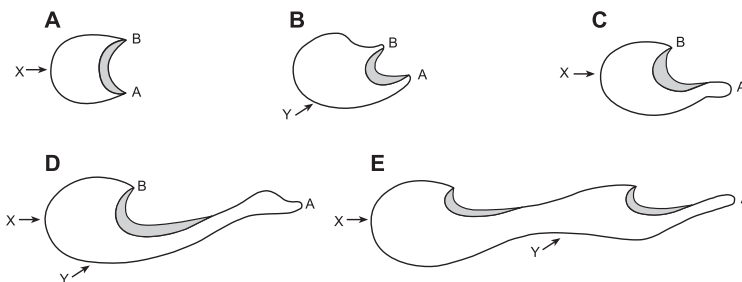


Figure 8.2. Progressive evolution of a barchan dune into a linear dune, showing associated sand-moving wind direction. (After Sparks, 1972, modified from Bagnold, 1941.)

the specialist texts (e.g., Pye and Tsoar, 1990; Pye and Lancaster, 1993; Lancaster, 1995; Tchakerian, 1995; Goudie et al., 1999; Warren, 2013) and the references listed at the beginning of this section.

8.2 World distribution of active and stable dunes

Perhaps the most interesting aspect of desert dunes from a paleoclimatic perspective is the present-day global distribution of active and stable dunes (Figures 8.3 and 8.4). For example, along the southern margins of the Sahara, a wide belt of now vegetated and stable dunes extends for at least 500 km to the south of the present-day southern limit of active dunes in the Sahara, which coincides with the 200 mm isohyet, presumably indicating greater-than-present aridity when those fixed dunes were active. Likewise, in the Thar Desert of north-west India, vegetated and now stable dunes extend to the east far beyond the present limit of mobile dunes into areas that receive well in excess of 450 mm of rain a year (Goudie et al., 1973). The critical questions here are when the fixed dunes south of the Sahara and the eastern Thar Desert were last active, and whether they were active at the same time. On a global basis, the area covered by active dunes during the last major phase of dune activity seems to show that the deserts were once far more extensive than they are today, at a time when the world climate was thought to be much drier in the intertropical zone (Sarnthein, 1978; Sarnthein et al., 1981). Indeed, Sarnthein (1978) commented quite explicitly that active sand dunes presently occupy about 10 per cent of the land area between 30°N and 30°S but that during the Last Glacial Maximum (21 ± 2 ka), the corresponding percentage of area covered by active dunes between those two latitudes was probably closer to 50 per cent, especially when taking into account the greater land area resulting from lower sea levels.

In order to test this model with appropriate rigour, two things are necessary. One is a robust chronology of dune formation, particularly because many linear dunes are polygenic (Fujioka et al., 2009; Cohen et al., 2010a; Singhvi et al., 2010; Fujioka and Chappell, 2011; Yang et al., 2011b). Difficulties arise in interpreting whether the ages obtained by luminescence dating of dune sand samples collected from different depths within the dune reflect peak dune activity or simply the final transitional phase from maximum movement to waning accumulation (Swezey, 2001; Swezey, 2003). The second requirement is a set of well-dated climate proxies that are able to show quite independently of the dune evidence whether the local or regional climate was indeed more arid than it is today in that region. Thomas (1997) has provided a comprehensive assessment of such evidence, which seems on balance to favour the glacial aridity model of Sarnthein (1978), at least for the hot tropical deserts, but the reality on a more local scale is in fact far more complex.

It is also not always obvious whether the inferred desert expansion involved a physical migration of dunes into what are now semi-arid and subhumid regions

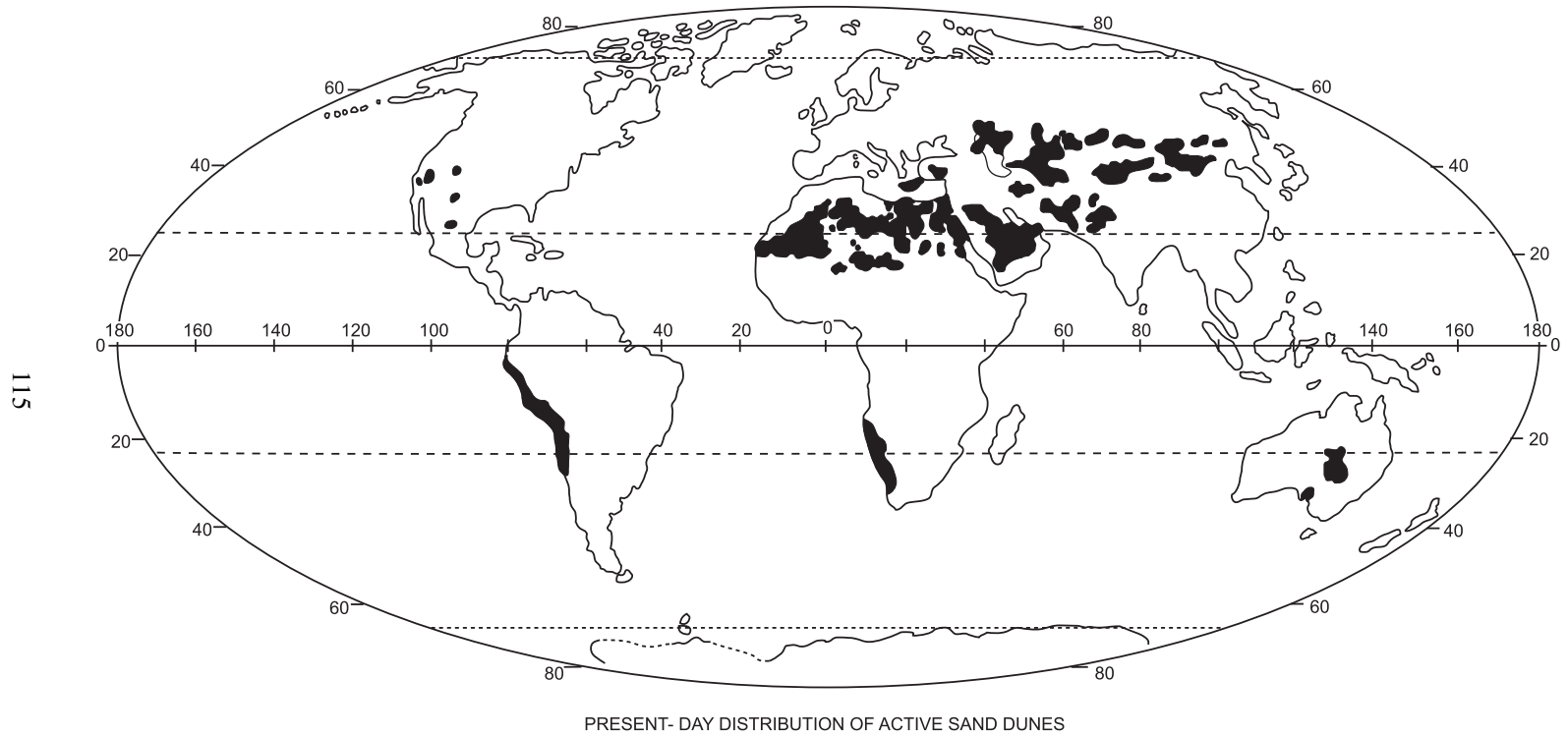
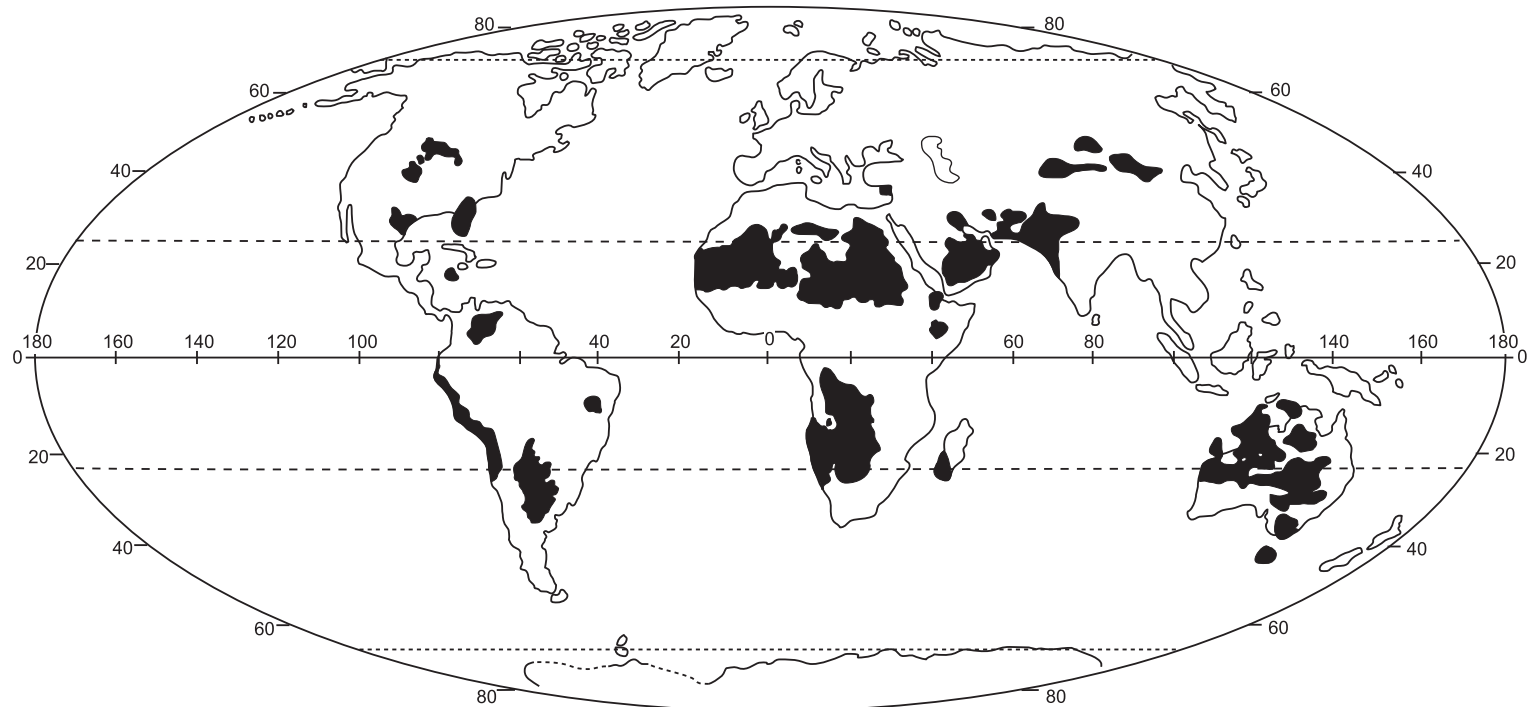


Figure 8.3. Map showing presently active desert dunes. (After Williams et al., 1998).



DISTRIBUTION OF ACTIVE SAND DUNES AT 18ka

Figure 8.4. Map showing desert dunes thought to have been active during the LGM. (After Williams et al., 1998, modified from Sarnthein, 1978.)

or simply a reactivation of pre-existing dunes and sand-plains that had lost their stabilising plant cover as a result of increased aridity or other reasons, including fire or the impact upon vegetation of changes in atmospheric carbon dioxide concentration (Hesse et al., 2005; Hesse, 2010). A fundamental issue that needs to be resolved is the unproven assumption that the presence of now vegetated and stable dunes in presently semi-arid areas does indeed reflect a former reduction in precipitation. As we shall see in the following sections, dune formation reflects a variety of controlling factors, each of which may obscure or outweigh the influence of local precipitation.

8.3 Sand transport by wind

Dune development is controlled by three main independent variables: wind speed, sand supply and vegetation cover. The early observations by Bagnold in the Libyan Desert, supplemented by his sand flume studies, demonstrated that the movement of individual sand grains by saltation is a function of wind velocity and grain size, with the threshold value for sand mobilisation and transport ranging from 4 m/sec (Bagnold, 1941, p. 70) to 6 m/sec, depending on sand particle size (Wasson et al., 1983, p. 126). As wind velocity increases, there is an exponential increase in sand movement (Figure 8.5). Analyses of changes in sand particle size in sediment cores off the west coast of the Sahara show that the Trade Winds were stronger during the last glacial, probably as a result of enhanced anticyclonic circulation over the tropical deserts linked to steeper temperature and pressure gradients between tropical and equatorial latitudes (Parkin and Shackleton, 1973; Parkin, 1974; Sarnthein et al., 1981).

Several quite different models of sand transport have been invoked to account for the formation and movement of linear dunes. One model favoured by Mabbutt (1977) and by Hollands et al. (2006) is the ‘wind-rift’ model in which turbulent wind vortices remove sand particles from the sand-floored swales between the linear dunes and deposit the sand on the flanks of the dunes. The parent sand in the swales can be alluvial, as in the case of the late Quaternary alluvial sands in the north-western Simpson Desert of Australia dated by Hollands et al. (2006), or reworked eolian sand formed during an earlier phase of dune degradation, as is presently occurring to linear dunes in the Kimberley region of north-west Australia (Goudie et al., 1993).

A frequently invoked model is one where there is movement by lateral accretion of sand in the downwind direction by sand avalanching at the proximal slip face, with the linear dune ultimately advancing several hundred kilometres or more from its initial point of origin (Twidale, 1972; Pye and Tsoar, 1990; Lancaster, 1995).

A third model involves vertical accretion (Telfer and Thomas, 2007; Stone and Thomas, 2008; Cohen et al., 2010a) and is supported by the fact that OSL ages become progressively older with depth within the same linear dune. This model seems to apply especially to linear dunes formed from transverse source-bordering dunes whose

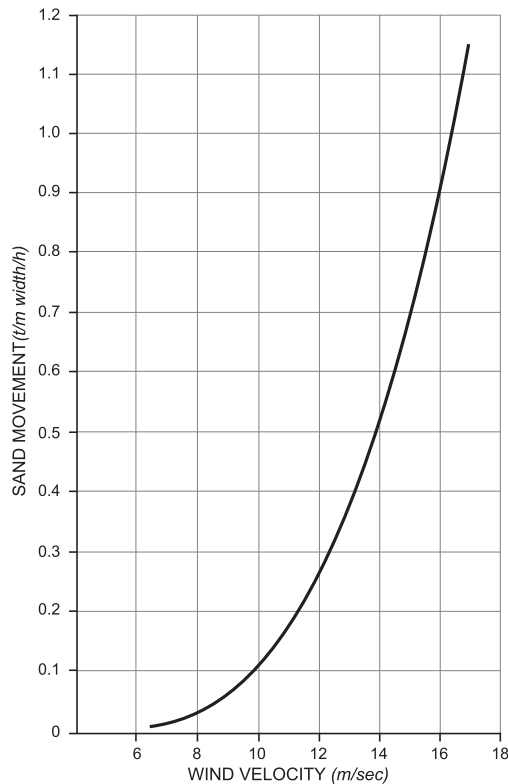


Figure 8.5. Wind velocity and sand movement. (After Williams et al., 1998, modified from Bagnold, 1941.)

origin was closely tied up with periodic influxes of fluvial sand. The last two models are, of course, not mutually exclusive, because some lateral accretion can also occur in dunes characterised by mainly vertical accretion.

8.4 Dune form and movement in relation to wind speed, sand supply and vegetation cover

As noted in Section 8.3, dune form varies in relation to wind speed, sand supply and vegetation cover. As a broad generalisation, transverse dunes (i.e., dunes oriented perpendicular to the dominant wind direction) are common in areas where winds are strong and sand supply is abundant, overriding the presence of any vegetation. They are therefore a poor indicator of aridity. Linear dunes are associated with moderately strong unidirectional or bidirectional winds, sparse to absent plant cover and moderate sand supply. Crescentic, or *barchan*, dunes are aligned with their horns oriented downwind (the reverse of transverse dunes), so the downwind slip face is at a right angle to the dominant wind direction. They can advance quite rapidly, at rates of up

to several metres a year (Haynes, 1989), under the influence of strong unidirectional winds and are common in areas of limited sand supply. Should they migrate into areas with a bidirectional wind regime, they may develop into linear dunes, depending on sand supply (Bagnold, 1941).

Luminescence dating of the linear dunes of the Simpson and Strzelecki deserts in central Australia shows that the crests have been repeatedly active to depths of many metres during the late Quaternary but appear virtually inactive today, a paradox noted by Hesse and Simpson (2006). These authors investigated the relationship between changes in vegetation cover and episodic sand movement on the dunes at three sites located along a climatic gradient in which the ratio of evaporation to precipitation increased from south to north. They surveyed the dune crests both during drought and after renewed rain. They concluded that ‘vegetation cover is very much a controlling factor in sand movement on Australian dunes and that most Australian dunes are inactive for most of the time because of the abundance of vegetation’ (Hesse and Simpson, 2006, p. 287).

Ash and Wasson (1983) also found an inverse relationship between plant cover and mobile sand and suggested that dunes become stable once the plant cover attains a threshold value of about 30 per cent. In contrast to Hesse and Simpson (2006), they concluded that sand movement is controlled by the frequency of strong winds. If strong winds are more frequent during times of drought, as is often the case, then both factors (reduced plant cover and high velocity winds) probably operate interactively, so it will not be easy to distinguish the relative importance of wind speed and plant cover as separate factors controlling sand movement. In addition, because the cover of annual grasses and forbs will be reduced during dry years, plant cover acts as an indirect index of precipitation and hence of aridity.

8.5 Dune orientation in relation to wind speed and direction

Linear dunes reflect the influence of the dominant sand-moving winds. In a pioneering study of the Holocene and older climates of the Sudan, Warren (1970) investigated the dominant sand-moving wind directions and the alignments of the vegetated dunes of Kordofan Province west of the White Nile. From the alignments of several generations of these now fixed linear dunes, he postulated a sequence of late Quaternary shifts in the wind and rainfall zones. In the absence of any dates, Warren tentatively considered the first and most arid phase to be late Pleistocene or older. He inferred a 450 km shift of both wind and rainfall belts to the south of their present position during that time. He regarded the wet phase that followed as equivalent to the terminal Pleistocene–early Holocene (15–8 ka) wet phase dated elsewhere along the southern margins of the Sahara. Warren estimated that during this time, there was a shift of the climatic and vegetation belts to roughly 250 km north of their present position, a conclusion accepted by Wickens (1975a; 1982) in his reconstruction of the hypothetical early



Figure 8.6. Source-bordering dune, lower Blue Nile, central Sudan.

Holocene vegetation of the Sudan. The second arid phase was brief and only involved a 200 km southward shift of the wind and rainfall belts. Warren equated the final moist phase with the moist Neolithic phase, dated to 7–5 ka in Chad. However, in the absence of any direct dating of the dunes themselves, these conclusions must remain speculative – working hypotheses to be tested by future investigation involving luminescence dating of the dune sands.

8.6 Paleochannels and source-bordering dunes

One particular type of dune common in areas where rivers flow into deserts is genetically associated with the presence of a regular supply of alluvial sand transported by seasonally flowing streams. Such dunes are termed source-bordering dunes (Figure 8.6) but should more strictly be defined as ‘fluvial’ or ‘riverine source-bordering sand dunes’ (Page et al., 2001), and they have attracted particular attention in Australia (Wasson, 1976; Bowler, 1978a; Bowler, 1978b; Williams et al., 1991a; Nanson et al., 1995; Page et al., 2001; Maroulis et al., 2007; Cohen et al., 2010a). Three conditions appear to be necessary for the formation of fluvial source-bordering dunes. The first prerequisite is a regular supply of bed-load sands brought in by rivers that dry out seasonally, leaving their sandy point-bars exposed to deflation. The second requirement is an absence of riparian vegetation so that sand movement out of the

channel through deflation is not impeded. The third condition is a regime of strong, unidirectional winds to allow the mobilisation and transport of sand from the dry river channel to form linear dunes downwind of the channel. In regard to the first condition, for the dunes to develop and continue to extend downwind, the alluvial sand supply needs to be regularly replenished. We discuss the relationship between Quaternary fluvial activity, climate and source-bordering dune formation in greater detail in [Section 8.15](#).

8.7 Lunettes and clay dunes

Lunettes are another form of source-bordering dune. However, they are invariably associated with desert lakes rather than river channels, and they occur as transverse dunes located on the downwind margin of playa lakes. They owe their original French name to their crescent shape, with the concave margin facing upwind, in contrast to barchan dunes in which the concave margin lies to leeward. Initially widely recognised in north-west Africa (Boulaine, [1954](#)), they are very common in the drier regions of south-east and south-west Australia, as well as the area adjoining the desert pans in southern Africa (Shaw and Thomas, [1989](#)). The lithology of lunettes can be highly variable, ranging from sand to clay to gypsum (Coque, [1962](#); Bowler, [1973](#); Benazzouz, [1986](#); Shaw and Thomas, [1989](#); Williams et al., [1991](#)). The clay is present in pelletal form, and the sand may contain wind-blown charcoal and even the calcareous oogonia of charophytes (Williams et al., [1991](#)).

Bowler ([1973](#)) proposed a simple model to account for polygenic lunettes, such as those characteristic of late Pleistocene Lake Mungo in semi-arid western New South Wales, in which a basal unit of sand is capped by finely laminated pelletal clay. The sand is blown from beach sands laid down during times of high lake level. Once the lake level drops, the fine-grained lake floor silts and clays become exposed to seasonal deflation during the hot dry summers, and the loose aggregates, formed as the saline mud dries out, are transported as sand-sized particles to form thin sheets of clay ([Figure 8.7](#)). Lunettes can therefore provide a detailed history of transgressions and regressions in their parent lake (Williams et al., [1991](#); Dutkiewicz and von der Borch, [1995](#); Bowler, [1998](#); Bowler and Price, [1998](#); Dutkiewicz and von der Borch, [2002](#)) and are especially amenable to dating using luminescence techniques (Dutkiewicz and Prescott, [1997](#); Dutkiewicz et al., [2002](#)).

8.8 Paleomagnetic and luminescence dating of eolian sands and silts

There have been a number of attempts to date times of dune activity using thermoluminescence (TL) (Stokes et al., [1997](#); Huntley and Prescott, [2001](#); Singhvi et al., [2010](#)) and, more recently, optically stimulated luminescence (OSL) dating methods (Stokes et al., [1998](#); Teeuw and Rhodes, [2004](#); Tchakerian, [2009](#); Williams et al., [2010b](#)).

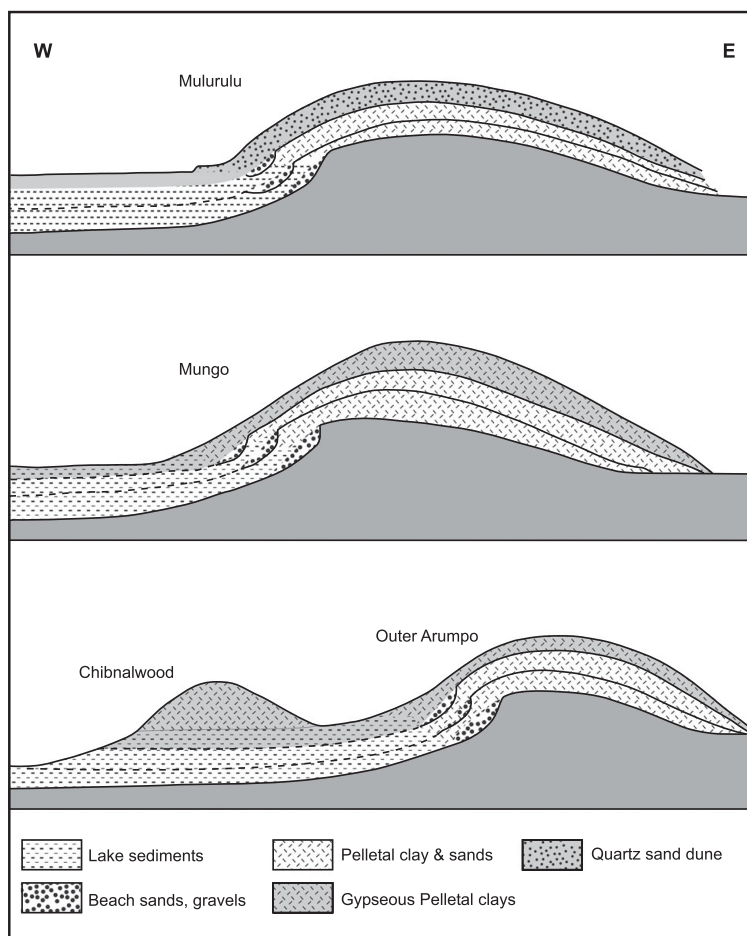


Figure 8.7. Model of a lunette formation as a lake dries out. (After Bowler et al., 2011.)

The luminescence dating method indicates the last time when the sand grains were exposed to daylight and the ‘luminescence clock’ was reset to zero, as discussed in [Chapter 6](#). Because most dune sands are devoid of organic remains, radiocarbon dating is not an option. In the case of deep playa lake sequences where dunes overlie playa clays, it is sometimes possible to use a combination of paleomagnetic and luminescence dating to determine long-term hydrological changes and the onset of desert dune formation (Chen, 1989; Chen et al., 1990; Chen and Barton, 1991; Chen et al., 1995).

Considerable caution is advisable when interpreting TL and OSL ages obtained on dunes. Depending on the sampling interval selected, one can be misled into believing there have been discrete phases of dune building (Stokes et al., 1997), when a closer sampling interval may denote almost continuous accretion. Furthermore,

the ages will only relate to sand that has been preserved at that site at that time, and so is not necessarily indicative of major episodes of dune activity, as discussed in [Section 8.14](#).

8.9 Problems in using desert dunes to reconstruct climatic change

Because active dunes are a feature of all deserts, it is natural to assume that the presence of now stable and vegetated dunes along the margins of many deserts indicates that these regions were formed at a time when the arid zone was more extensive than it is today. In addition, where cores have been collected from linear dunes for TL or OSL dating, it is again often assumed somewhat uncritically that the ages obtained reflect intervals of sand deposition and dune activity associated with greater aridity. Because dune formation and movement is a complex function of effective precipitation, wind velocity, sand supply and surface plant cover, it is instructive to examine a number of case histories from different deserts to test whether or not dune activity is indeed a good indicator of former aridity. The results of this exercise will prove surprising. Our aim is not to provide a comprehensive review of dune studies on each continent but to adopt a more selective approach and consider the scope and limitations of using desert dunes to reconstruct past climates. We begin with the Northern Hemisphere deserts of the Sahara, Negev and Sinai, Arabia, Pakistan, India and northern China before moving to the Southern Hemisphere deserts of the Namib, Kalahari and Australia.

8.10 Desert dunes of the Sahara

During the Neogene, the central Sahara was drained by a series of major river systems that flowed from the northern Chad Basin east of Tibesti across the Libyan Desert to flow into the Mediterranean (Griffin, [1999](#); Griffin, [2002](#); Griffin, [2006](#); Griffin, [2011](#)). These rivers dried out in the late Miocene to early Pliocene as the Sahara became progressively more arid (Griffin, [2002](#); Griffin, [2011](#)). A series of very large, sinuous, dry river valleys is a legacy of this time and is clearly visible on satellite images. Aridity was accentuated by tectonic uplift in East Africa that caused a major change in atmospheric circulation and led to a reduction in rainfall over the Chad Basin (Sepulchre et al., [2006](#)). The change in rainfall regime over East Africa resulted in a change from tropical forest to open grassland and woodland and was associated with the proliferation of the Pliocene hominids unique to Africa (Williams et al., [1998](#); Sepulchre et al., [2006](#); Cerling et al., [2011](#)).

Miocene uplift led to an acceleration of erosion in the Saharan uplands. Much of the resulting sediment was carried to the sea by big rivers like the Nile, Niger and Senegal, but a considerable amount began to accumulate in large subsiding sedimentary basins such as the Kufra-Sirte Basin in Libya and the Chad Basin, providing the source material for the Quaternary and possibly older desert dunes. In the Chad Basin,

Servant (1973) identified wind-blown sands in a number of what he considered to be very late Tertiary stratigraphic sections. He concluded that the onset of aridity and the first appearance of desert dunes in this part of the southern Sahara was a late Tertiary phenomenon. Using fossil and sedimentary evidence, Schuster et al. (2006) have since confirmed that the onset of recurrent desert conditions in the Chad Basin began at least 7 Ma ago. The sedimentological evidence of Servant (1973) and of Schuster et al. (2006) and the pollen evidence of Maley (1980; 1981; 1996), indicate that the onset of climatic desiccation and the ensuing disruption of the integrated early to mid-Miocene Saharan drainage network (Griffin, 2006) was a feature of the late Miocene.

The major dune fields of the Sahara occupy topographic depressions formed during the Miocene and earlier phases of volcanism and tectonic activity in the central, eastern and northern Sahara. It was during this time that volcanic massifs such as the Hoggar, Tibesti and Jebel Marra were formed. Volcanism was preceded and accompanied by uplift of the Proterozoic crystalline basement complex rocks and their sedimentary cover of Palaeozoic and Mesozoic sandstones and shales, giving rise to the undulating topography of basins and swells described by Arthur Holmes nearly fifty years ago (Holmes, 1965, fig. 763). Eocene deep weathering was followed by climatic desiccation and by tectonic uplift, triggering a wave of erosion. As the once well-integrated drainage network of the Sahara became segmented and disorganised, the alluvial sediments were no longer carried to the sea and instead accumulated within internally drained depressions, or depocentres. The finer particles were winnowed out by the prevailing Trade Winds along the southern Sahara and blown out across the Atlantic as desert dust, bringing nutrients to the Amazon rainforest (see Chapter 9). In the north of the Sahara, the westerlies carried Saharan dust at least as far as the Negev Desert in southern Israel. The sand-sized particles left behind by this winnowing process were in turn fashioned into desert dunes and sand plains.

As a very rough rule of thumb, the orientation of the dunes reflects the anticyclonic wind circulation in the Sahara, but in contrast to central Australia, the pattern is far from simple and there is considerable evidence of the deflection of wind and dunes around mountains and other smaller topographic obstacles (Grove, 1980; Mainguet et al., 1980; Warren, 2013). The age of such major dune fields as the Grand Erg Oriental (Great Eastern Erg, or sand sea) in southern Algeria is not well-established but appears to be at least as old as the fluvial deposits that traverse them, some of which contain Early Stone Age hand-axes (Acheulian bifaces) with a minimum age of 300 ka and a maximum age of approximately 1.5 Ma (see Chapter 17). It is still an open question as to whether there has been much long-distance movement of sand across the Sahara or whether certain dune fields are essentially local features, as seems to be the case with the Qoz Dango in south-west Darfur (Williams et al., 1980).

Along the southern margins of the Sahara, as mentioned in Section 8.2, presently fixed dunes extend to at least 500 km south of the present limit of active dunes, which

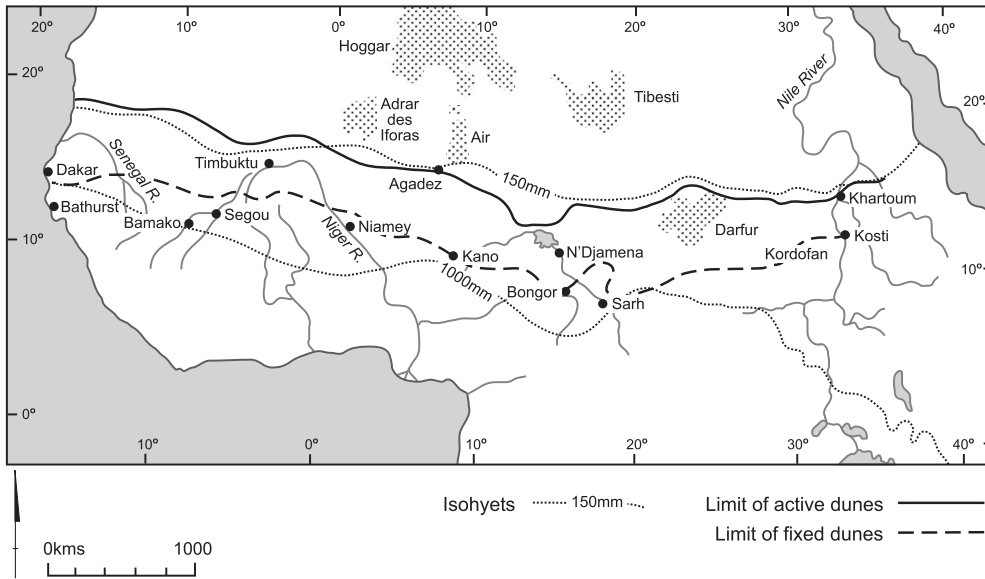


Figure 8.8. Map showing active and fixed dunes in and beyond the Sahara. The present-day limit of active dunes is bounded by the 150 mm isohyet. Fixed dunes extend up to 500 km south of the Sahara, locally into areas that now receive 1,000 mm of mean annual rainfall. (Modified from Williams et al. 1998, fig. 9.8.)

today is broadly delineated by the 200 mm isohyet (Figure 8.8) (Grove, 1958; Grove and Warren, 1968; Mainguet and Canon, 1976; Mainguet et al., 1980; Talbot, 1980). Based on an array of ^{14}C ages of fluvial and/or lacustrine sediments overlying and underlying the fixed dunes, most workers concluded that the time of peak dune activity coincided with the cold, dry and windy Last Glacial Maximum, some 20,000 years ago. This was also a time when the great lakes of East Africa dried up or became saline (see Chapters 11 and 12) and when exports of desert dust from the Sahara were exceptionally high (see Chapter 9). However, the reality may be somewhat more complex.

Based on his Tunisian experience, Swezey (2001; 2003) has argued that dune sediments are best preserved if succeeded by a humid phase but have less chance of preservation if followed by an arid phase. We noted in Section 8.3 that glacial age wind velocities were stronger in the Sahara than they are today (Lancaster et al., 2002), an inference confirmed by the abundance of desert dust in marine cores off the west Saharan coast (deMenocal et al., 2000). However, in the most comprehensive review of dune activity in the Sahara yet attempted, Swezey (2001) determined that the vast majority of dated eolian records from the Sahara were in fact younger than the Last Glacial Maximum (21 ± 2 ka). He concluded that this apparent absence of evidence for LGM dune activity was probably caused by subsequent reworking during the arid Younger Dryas episode (12.8–11.5 ka) and made the important point that this portion

of the record had been preserved as a result of a change to wetter conditions in the Sahara between 11.5 and 7 ka.

8.11 Desert dunes of the Sinai and Negev deserts

Efforts to extract information about past climatic changes from desert dunes can often lead to unexpected and sometimes counter-intuitive results, as shown by recent studies from the Negev Desert of Israel (Roskin et al., 2011a; Roskin et al., 2011b). Vegetated and stabilised linear dunes occupy a dune field of 1,300 km² in area in the north-west Negev Desert and form the eastern extremity of the northern Sinai sand sea. The main phase of dune accumulation began as recently as 23 ka, although there had been sporadic dune deposition since about 100 ka (Roskin et al., 2011a). A detailed set of 97 OSL ages obtained from thirty-five dunes and inter-dune swales point to three main episodes of dune mobility: 18–11.5 ka (post-LGM), 2–0.8 ka (very late Holocene) and the last 150 years (modern). The post-LGM interval was the most widespread phase of dune movement, and it involved dune damming of certain small valleys, which led to impeded drainage and the formation of small lakes and ponds between the dunes.

Late Pleistocene lowering of the sea level by about 120 m led to the exposure of the previously submerged sands in the Nile Delta. These sands were carried eastwards by longshore drift and blown inland and eastwards to feed the sand dunes of the Sinai Desert and northern Negev. Initial advance of the linear dunes from the Sinai into the northern Negev was underway during the LGM (23–18 ka), but gained momentum around 16–13.7 ka, with a later minor phase of advance at 12.4–11.6 ka (Roskin et al., 2011b). These two phases were synchronous, respectively, with Heinrich Event 1 in the North Atlantic and with the Younger Dryas cold event in Greenland.

The orientation of the linear dunes indicates that the sand-moving winds were blowing from the west. Roskin et al. (2011b) concluded that movement of the linear dunes during the late Pleistocene was associated with stormy winter cyclones from the eastern Mediterranean. Such cyclones would have brought both violent winds and more rainfall, indicated by the presence of lakes, paleosols and prehistoric occupation sites in the swales between the linear dunes. Although the Holocene climate was more arid than the climate that prevailed during the late Pleistocene, the decrease in storminess led to dune stabilisation. Wind regime was thus more important than aridity in controlling the movement and eventual stabilisation of the vegetated linear dunes in the northern Negev Desert. We thus have the paradox of maximum dune activity coinciding with a time of increased and not decreased rainfall in this region.

8.12 Desert dunes of Arabia, Pakistan and India

Two dominant wind systems control the distribution and orientation of the sand dunes in the deserts of peninsular Arabia (Glennie et al., 2002). The *Shamal* (Arabic for

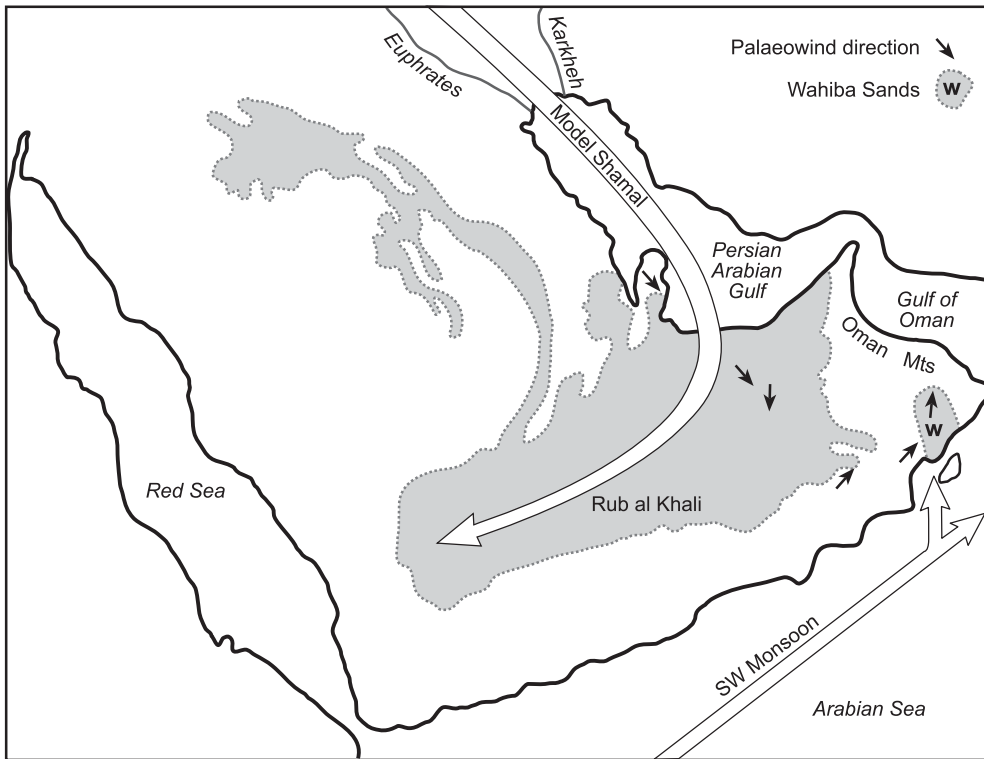


Figure 8.9. Map showing the dominant wind systems in the Arabian Peninsula and the location of the Rub al Khali and the Wahiba Sands. (After Glennie et al., 2002 and Singhvi et al., 2012.)

‘north’) is a strong wind that blows from the north-west down the Persian/Arabian Gulf before swinging to the south-west and across the hyper-arid Rub al Khali Desert (Figure 8.9). This wind is most evident in the cold winter months and is reminiscent of the cold north wind that blows up the Nile Valley in winter, causing sand storms and low visibility almost as far south as Khartoum, as in January 2012. The other strong wind system is that of the south-west monsoon, which fashioned the dunes of the Wahiba Sands in eastern Oman, as well as the south-west- to north-east-aligned linear dunes of the Thar Desert in north-west India. A limited number of TL, OSL and ^{14}C ages obtained from sites scattered across a huge area suggest that the history of dune activity in the far south-east of the Arabian Peninsula, notably in the Wahiba Sands, differs from that inferred for the rest of the region and is similar to that now evident in the Thar Desert. As in North Africa, the Last Glacial Maximum (21 ± 2 ka) was especially cold, dry and windy in the Rub al Khali and adjacent sand deserts of Arabia, but the most recent maximum dune accretion occurred somewhat later in the Wahiba Sands and in the Thar Desert and took place at about the time that the south-west summer monsoon was becoming stronger once more, in the very late

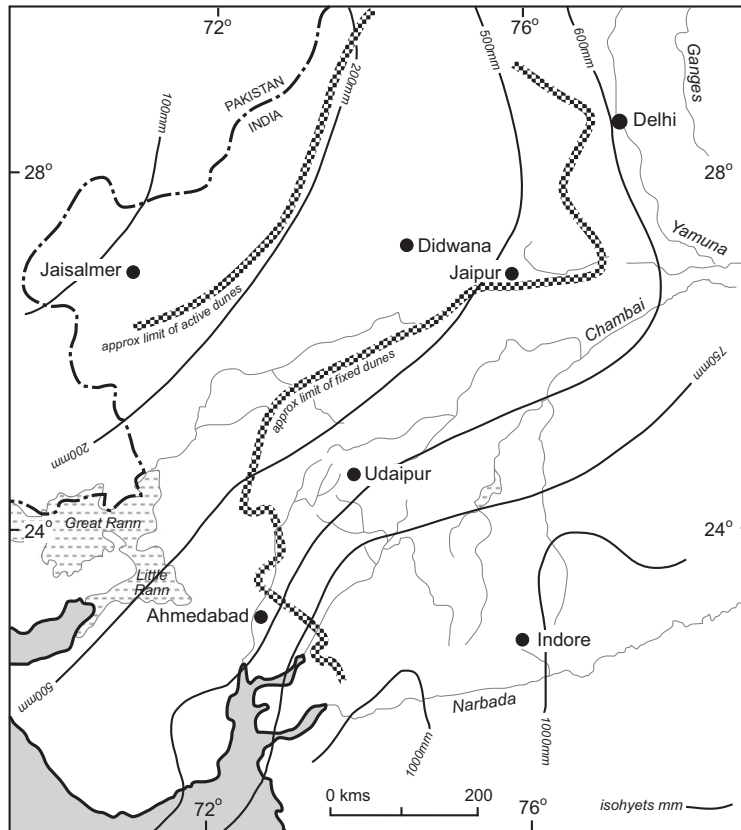


Figure 8.10. Map showing active and fixed dunes in the Thar Desert, India. Active dunes occur where the mean annual rainfall is less than 250 mm; fixed dunes extend into areas that now receive at least 450 mm of mean annual rainfall. (After Singhvi et al., 2010.)

Pleistocene, about 15–14 ka (Wasson et al., 1983; Chawla et al., 1992; Dhir et al., 1992; Thomas et al., 1999; Singhvi et al., 2010).

The Thar Desert occupies an area of roughly 320,000 km² in north-west India and eastern Pakistan between latitudes 24°30'N and 30°N and 69°30'E and 76°E. It extends roughly 800 km (500 miles) from WSW to ENE and 400 km (250 miles) from north-west to south-east (Figure 8.10), and occupies much of the Indian State of Rajasthan and the Pakistan Province of Sind. Mean annual rainfall decreases from more than 500 mm in the east to less than 100 mm in the west, near the edge of the Indus Valley. The Aravalli Hills form an approximate eastern boundary to the Thar Desert and are the source of a network of seasonal and ephemeral rivers that flow westwards into the desert, including the one integrated drainage system in this desert – the seasonal Luni River. The Thar Desert is flanked to the west by the Cholistan Desert of eastern Pakistan, most of which is situated within Sind Province, to the

south by the salt marshes of the Rann of Kutch and to the north by the presently dry valley of the Ghaggar (see [Chapter 10](#)). The Thar Desert forms the eastern terminus of the wide stretch of tropical deserts that extend from the Sahara across Arabia, Iraq, Iran, Afghanistan and Pakistan to north-west India, a distance of 8,000 km, spanning 110° of longitude. It therefore lies within the tropical northern Eremian vegetation zone, comprising the Saharo-Sindian flora and the Irano-Touranian flora.

Much of the better-watered eastern desert is well-vegetated, and the flora is remarkably reminiscent of that found along the southern Sahara and across Arabia, including such familiar trees as *Acacia senegal*, *Salvadora* spp., *Tamarix* spp., *Calotropis procera*, *Ziziphus* spp. and *Leptadenia pyrotechnica*, as well as the well-known dune stabilising grasses *Panicum turgidum*, *Cenchrus biflorus* and *Eragrostis* spp., with the ubiquitous bitter melon *Citrullus colocynthis* trailing along the ground.

We saw in [Section 8.2](#) that in the Thar Desert, vegetated dunes extend well to the east of the present desert into areas with a mean annual rainfall of 450 mm or more. Leaving aside the question of human disturbance, active dunes in this desert are today mainly confined to areas with less than 250 mm annual rainfall. Goudie et al. (1973) considered that this implied that the climate had been drier and the desert more extensive during the very late Pleistocene. Wasson et al. (1983) questioned this interpretation on the grounds that even vegetated dunes could be mobile and, in any event, dune sands continued to accumulate well into the Holocene.

We noted earlier in this section that the Thar Desert dunes reflect the influence of the south-west summer monsoon winds, with the linear dunes in the drier western half of the desert oriented parallel to the dominant sand-transporting wind direction. In favourable circumstances, such as in the lee of ranges of low hills aligned roughly perpendicular to the dominant sand-moving winds, a long, nearly continuous record of dune accretion may be preserved. One such dune, near Didwana in the north of the desert, with the unromantic site name of 16R, contains a remarkably well-preserved sequence of twelve cycles of dune accretion, soil formation, calcrete development and subsequent erosion spanning the last 190 ka ([Figure 8.11](#)) (Singhvi et al., 2010). It also contains a stratified sequence of prehistoric stone artefacts ranging from Lower through Middle and Upper Palaeolithic to Mesolithic in age, synchronous with more humid climatic interludes in this region (Misra, 1983; Dhir et al., 1992; Dhir et al., 2010; Singhvi et al., 2012). The calculated time interval between successive phases of dune sand accumulation ranged from 22.2 ka to 15.8 ka, with a mean of 19.0 ka. These values are consistent with a precessional influence (see [Chapter 3](#)) on dune activity. Initial dune accretion was associated with the onset of early monsoonal activity in this region. Carbon isotopes measured on organic matter within the sand profiles show consistent values close to $-21.6 \pm 1\%$, pointing to deposition during a transitional climatic regime characterised by a change from open C₃ grassland to C₄ woodland or forest (Singhvi et al., 2010).

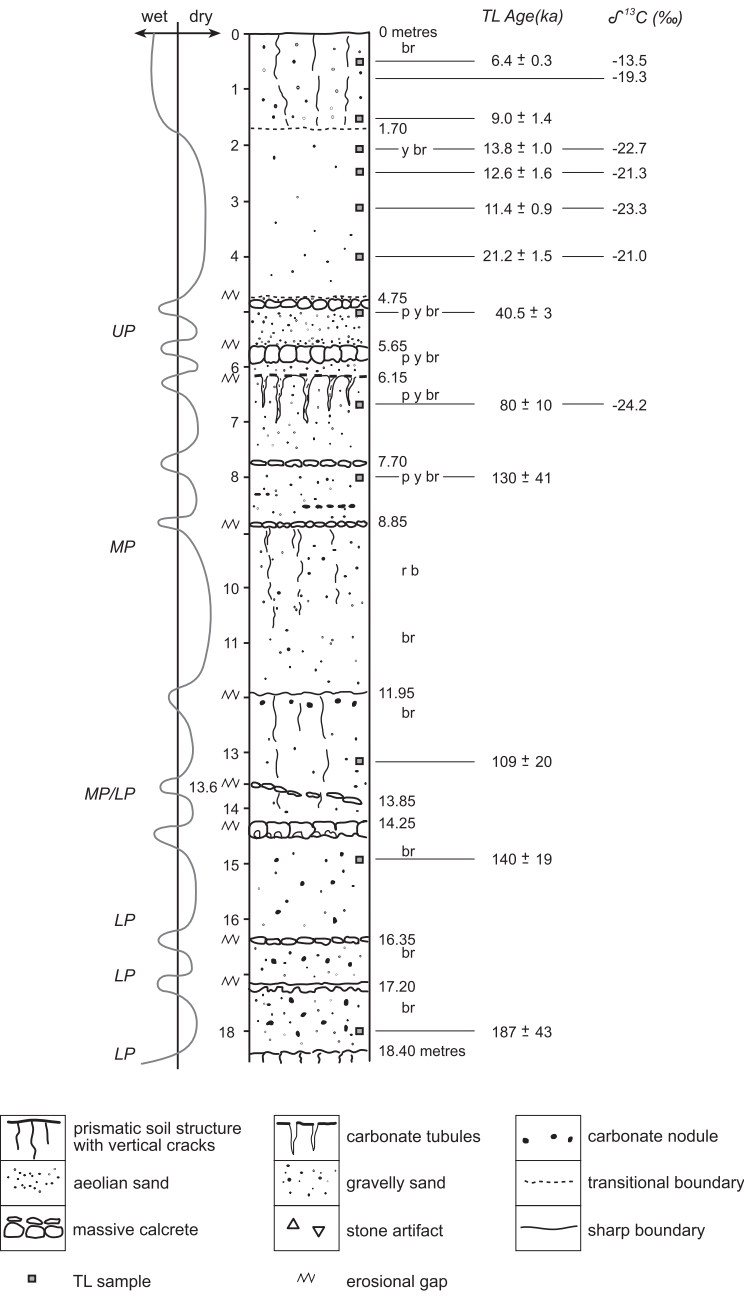


Figure 8.11. Stratigraphic section through a Quaternary polygenic dune in the Thar Desert, India, showing eleven alternating phases of soil/calcrete formation and sand accretion during the last 200,000 years. (After Singhvi et al., 2010.)

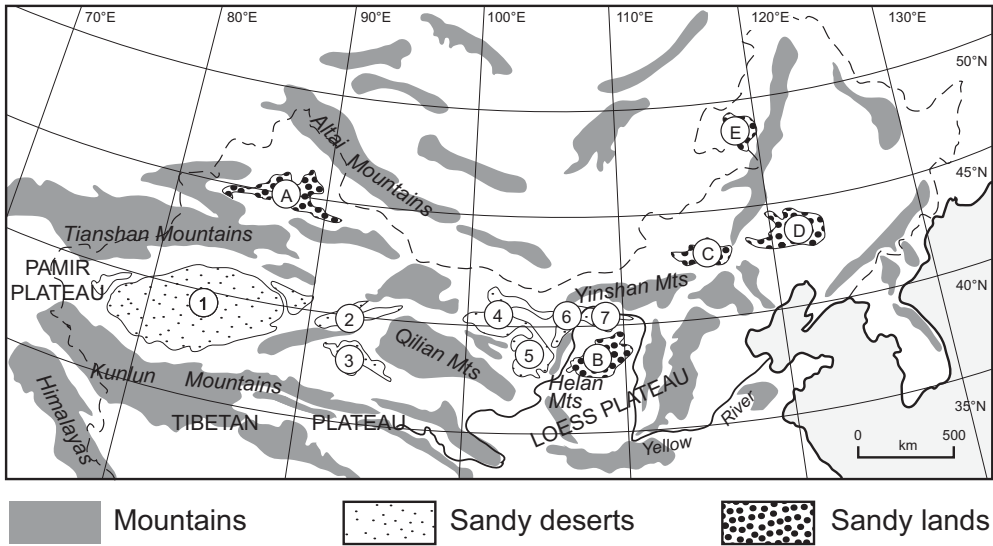


Figure 8.12. Map showing sandy deserts (active dune fields: 1 to 7) and sandy lands (areas of stabilised dunes: A to E) in northern China. (After Yang et al., 2011b.) Major mountain ranges in grey. 1. Taklamakan Desert; 4. Badain Jaran Desert; 5. Tengger Desert.

During drier intervals in the late Pleistocene, the coastal dunes in the tropical south of the subcontinent were also active (Jayangondaperumal et al., 2012). These dunes are also polygenic and consist of alternating fossil soils and wind-blown sand, as in the Thar Desert dunes.

8.13 Desert dunes of northern China

There is a considerable body of descriptive empirical research dealing with the desert dunes in China. Much of this work had a strong practical focus and was linked to efforts to prevent the movement of mobile dunes across certain strategically important roads and railways in arid northern and western China. Recent discoveries of oil in the Taklamakan Desert (Figure 8.12) have likewise encouraged dune stabilisation measures along key access roads during the past twenty years and have drawn on studies of dune movement carried out by the late Professor Zhu Zhenda and his colleagues during his time as Director of the Chinese Academy of Sciences (CAS) Desert Research Institute at Lanzhou, located in the northern Loess Plateau (Zhu et al., 1989; Zhu and Wang, 1992; Zhu and Wang, 1993). The detailed maps of the various dune fields produced during the course of this work remain useful to this day and provide valuable information on rates of dune movement in relation to dune type.

In contrast to India, discussed in the previous section, which has one large sandy desert only, China has twelve distinct deserts or sandy lands, only some of which have

yet been studied in any real detail from a perspective of past climatic changes (Yang, 1991; Derbyshire and Goudie, 1997; Yang, 2002; Yang and Scuderi, 2010; Yang et al., 2011a; Yang et al., 2011b; Yang et al., 2012). The Chinese deserts differ from their tropical Saharan, Arabian and Australian counterparts in three important respects. First, they are located in mid-latitudes rather than tropical latitudes and therefore lie within the zone of the mid-latitude westerlies rather than being located within the zone of dry subsiding air masses associated with the tropical anticyclones discussed in Chapter 2. They owe their aridity to distance inland and to pronounced rain shadow effects. Second, the deserts west of the Helan Shan (*shan* means mountain range) occupy tectonic depressions bounded by high mountain ranges (Figure 8.12). The higher of these mountains have permanent snow and ice, meltwater from which feeds rivers that are often substantial and flow into and, on occasions in the past, across the present dune fields. Location within large and deep tectonic depressions means that one important source of sediment for the evolving sand seas comes from the large alluvial fans flanking the foot of the mountains and from what Weissmann et al. (2010) termed ‘distributive fluvial systems’, which encompass the ‘mega-fans’ identified by Leier et al. (2005) that are discussed in Chapter 10. Third, and unsurprisingly, given the active tectonic history of western China (and central Asia more generally), the Chinese deserts span a considerable altitudinal range. The dune fields range in elevation from less than 1,000 m to nearly 5,000 m on the Tibetan Plateau and appear to occur at discrete elevations (Jäkel, 2002), although this may be coincidental.

In China, the term ‘desert’ (*shamo*) is restricted to active sand seas, and the term ‘sandy land’ denotes areas of fixed and vegetated sand dunes. According to this well-established classification, there are seven sand sea deserts in China and five main areas of sandy land (Figure 8.12). These sandy lands are the Chinese equivalents of the ancient erg of Hausaland in northern Nigeria described by Grove (1958) and the vegetated dune fields of the northern Kalahari (Thomas, 1984). As a general rule, the desert sand seas of northern China are confined to the most arid areas, while the sandy lands lie within the semi-arid regions of the centre and east (Figure 8.12).

Although Yang et al. (2011a) concluded from their comprehensive review of Quaternary environments in the desert lands of northern China that ‘little is also known about why dune type and size vary greatly across the drylands of China and the forcing factors that cause these differences’, there is still a great deal of useful environmental information that can be gleaned from the dunes within this vast region. As with any scientific research, the interpretation of this information remains a work in progress.

The Taklamakan is the largest desert in China, with an area of 337,600 km². Active dunes up to and slightly more than 100 m in height cover 80–85 per cent of its area. The setting of this desert within the almond-shaped Tarim Basin is spectacular, with the snow-covered Tian Shan forming the northern boundary and the similarly snow-covered Kunlun Shan forming the southern boundary (Figure 8.12). One or more vast lakes seem to have occupied the centre of the basin to an elevation of at least 1,100 m

during the early Pleistocene. As the lake(s) shrank progressively (or dried out and refilled), dunes developed along their sandy shorelines no doubt from sands ferried in by rivers from the adjacent high mountains. Over time, much of the evidence of these former vast lakes has either been buried or eroded (Yang et al., 2011b). It thus appears that the present sand sea in the Taklamakan is a relatively youthful feature, associated with a progressively more arid climate, although we cannot as yet provide any reliable suggestions as to when it first formed. During the late Pleistocene and mid-Holocene, rivers flowed through the dunes, and small lakes came into being, some as recently as three centuries ago (Yang, 2001a). However, a combination of climatic desiccation and human deforestation has hastened the processes of desertification in this region (Yang et al., 2011b).

The Badain Jaran Desert is the second largest desert in China, with an area of 49,200 km², and has been studied in some detail by Professor Yang Xiaoping and his colleagues (Yang, 1991; Yang and Williams, 2003; Yang et al., 2010; Yang et al., 2011a; Yang et al., 2012). This desert is bounded to the south by the glaciated Qilian Shan and adjoins the Tengger Desert (42,700 km²) to the east, with the north-south aligned Helan Shan forming the rugged eastern boundary and a convenient zone separating arid areas to the west and semi-arid areas to the east. The southern Badain Jaran Desert is renowned for the height of its giant dunes, which are mostly 200 to 300 m high but attain a maximum height of 460 m, making them not only the highest dunes on this planet, but also higher than the dunes on Mars (Yang et al., 2011a). A combination of factors seems to be responsible for the great height of these dunes, including a hilly subsurface bedrock topography, an abundance of fluvial sands, a complex wind regime and periodic stabilisation of the dune surface by calcareous soils during wetter climatic intervals.

This desert is also remarkable for the very large number of lakes – well over a hundred – that occupy the swales between the dunes and provide water and fodder for the Mongolian herders and their flocks of two-humped Bactrian camels during the summer months (Yang, 1991; Yang and Williams, 2003; Yang et al., 2010). We discuss the hydrological and climatic significance of these lakes in [Chapter 11](#).

A much-debated point is whether the different sand seas and sandy lands have been in contact with each other during past phases of active sand movement. Recent studies (summarised by Yang et al., 2012) of the particle-size distribution, heavy mineral content and quartz grain isotopic geochemistry in the various deserts and sandy lands have demonstrated reasonably convincingly that each desert operated as a self-contained unit, receiving its sand supply from river systems flowing from the mountains adjoining the particular deserts.

Where lake sediments either underlie dune sands or are banked against dunes, a relative chronology of wetter and drier phases can be attempted, buttressed by OSL and radiocarbon dating, but variable and quite large ¹⁴C reservoir effects (see [Chapter 6](#)), sometimes up to several thousand years (Hofmann and Geyh, 1998), have

impeded efforts to establish a coherent chronology of past climatic fluctuations. The thoughtful review by Yang et al. (2011b) enlarges on these difficulties. Even when the dunes contain pedogenic carbonates sandwiched between dune sands, indicative of episodically wetter conditions, as in the Badain Jaran Desert, there are well-known problems in obtaining reliable radiocarbon ages for pedogenic carbonates (see Chapter 6). One viable but very time-consuming option is to date the quartz grains enclosed within the carbonate horizons using OSL (Singhvi et al., 1996). In the sandy lands in the wetter eastern semi-deserts, the fossil soils within the now stable dunes are red kraznozems (Yang et al., 2011b), which are comparatively rich in clay and, presumably, contain some organic carbon (see Chapter 15). It should therefore be possible to obtain a reliable AMS ^{14}C chronology for these paleosols in the future.

8.14 Desert dunes of the Namib and Kalahari

The Namib Desert covers 34,000 km² and extends from south to north for about 2,000 km along the west coast of southern Africa between 23°S and 28°S. The present-day sand sea is underlain by the Tsondab Sandstone Formation (TSF), a fossilised sand sea of pre-late Miocene (Ward, 1988) or Oligocene age (Besler, 1991). The TSF was eroded to a gently undulating surface during the late Miocene. Pliocene uplift and fluvial incision coincided with the development of the modern sand sea. The Kalahari Sands were mapped by Cooke (1958) and occupy a vast area extending from beyond the Congo River in the north to the Orange River in the south (Figure 8.13). Dingle et al. (1983, p. 293) suggested a tentative age of Mio-Pliocene for the oldest of the Kalahari dunes. If all of the Kalahari Sands are indeed eolian, then they are almost certainly diachronous in age, with the oldest sands forming first in the north as southern Africa drifted northwards into dry tropical latitudes during the early to middle Cenozoic, a topic discussed at greater length in Chapter 18.

Besler (1983; 1991) and Lancaster (1989) have studied the morphology of the Namib dunes and the processes responsible for their formation. However, relatively few of the inland dunes have been dated by OSL. Lancaster and Teller (1988) have described the four main types of inter-dune deposits in the Namib, comprising of coarse, poorly sorted eolian sands, calcareous lake sediments, coastal salt marshes and alluvial silts, but the precise ages of these deposits are still unknown.

Chase and Thomas (2007) obtained OSL ages for thirty-five sand samples from six dune cores along a 300 km transect close to the coast, extending from Cape Town in the south to the Namib Sand Sea in the north. They found five distinct peaks suggestive of sand dune activity at 73–63, 49–43, 33–30, 24–16 and 5–4 ka, and concluded that sand movement was more closely linked to wind strength and sand supply than to periods of increased aridity. Earlier work by O'Connor and Thomas (1999) on late Quaternary degraded linear dunes in western Zambia also concluded that sediment supply, in this case from the Zambezi River, had played a pivotal role in dune development in this region. However, they also considered that a reduction

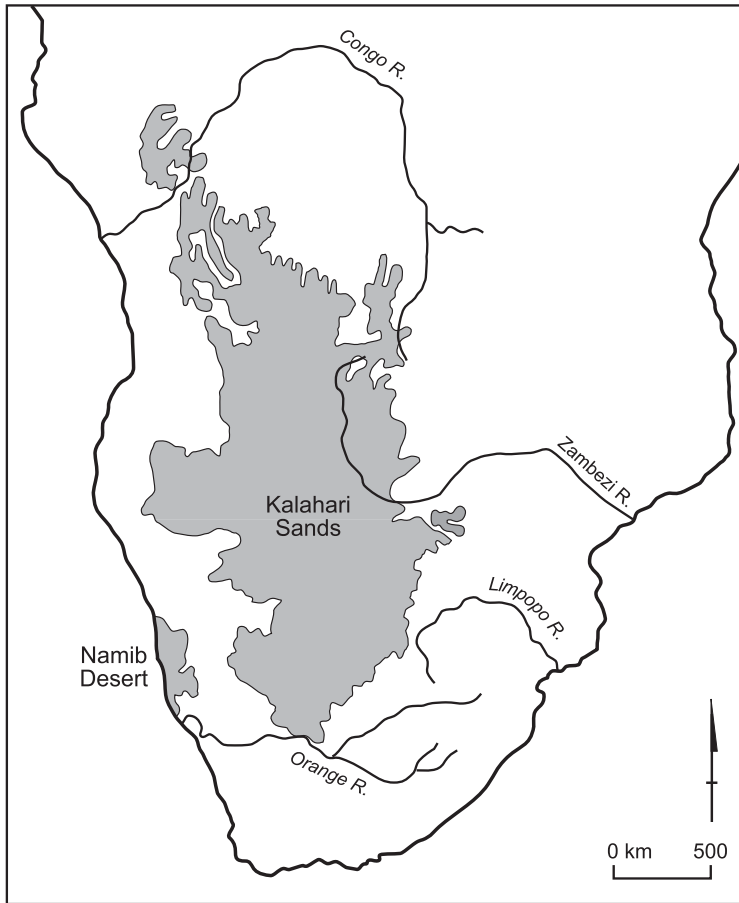


Figure 8.13. Map of the late Cenozoic Kalahari Sands. (After Cooke, 1958.)

in effective precipitation was necessary for dune formation and, most probably, an increase in windiness.

The linear dunes of the Kalahari are the most comprehensively dated dunes in southern Africa, but the interpretation of the luminescence ages has been a thorny problem. Initial work by Stokes et al. (1997; 1998) appeared to indicate multiple episodes of dune construction and inferred aridity in southern Africa during the last interglacial-glacial cycle at 115–95, 46–40, 26–20 and post-20 ka. The depositional gaps were considered to correspond to wetter climatic phases. These climatic fluctuations were thought to be associated with changes in sea surface temperatures in the south-east Atlantic and Indian Oceans, leading to changes in temperature gradients and movement of moist air masses into southern Africa from the north-east (Stokes et al., 1997; Stokes et al., 1998).

This reconnaissance style of dating was followed by a programme of detailed OSL dating of linear dunes at Witpan in the south-western Kalahari that showed initial

sand accumulation at 104 ka and further evidence of dune activity in spatially discrete locations at 77–76, 57–52, 35–27, 21–19 ka and again at 15–9 ka, when the climate in this region was unusually arid and linked to intensification of the continental anticyclone (Telfer and Thomas, 2007). Additional OSL dating, giving a total of 136 OSL ages for the southern Kalahari, including additional samples collected at closer vertical intervals of 0.5 m, suggested that the dunes in the south-western Kalahari had been partially active throughout much of the past 120 ka and that apparent age clusters could be produced spuriously as a function of reducing the sampling frequency with depth within the dunes (Stone and Thomas, 2008). Stone and Thomas (2008) concluded that the dunes in this region had been close to their threshold of reactivation throughout much of the late Quaternary and that earlier work invoking discrete phases of dune sand accumulation needed to be reassessed.

Chase (2009) reviewed the evidence for dune activity across southern Africa and concluded that three primary phases of dune activity centred on 60–40, 35–20 and 17–4 ka could be identified. Given the error terms associated with these ages, one could argue equally plausibly for more or less continuous dune activity from 60 ka onwards. Chase (2009) discounted aridity as the sole or even primary control over sand movement and invoked changes in wind strength as the main forcing factor. Chase and Brewer (2009) compared the output of coupled Ocean-Atmosphere Global Circulation Models with empirical studies of potential sand transport during the LGM (24–18 ka) and found little theoretical evidence for LGM dune mobility across southern Africa, in opposition to actual dated evidence of dune activity at this time obtained from field studies. They concluded that the model outputs could be unreliable, as could the empirical indices for dune activity. They also questioned the validity of using dune records as paleoclimatic proxies at millennial scales.

The evidence from desert dunes in southern Africa has yielded highly equivocal information about the Quaternary climates in this region, with some workers arguing that sediment supply and wind strength are more powerful determinants of dune activity than increased aridity. In addition, the search for pulses in dune accretion has been bedevilled by the conflicting results arising from differences in the sediment sampling resolution in regard to depth, with apparent pulses vanishing once dunes were sampled at closer vertical intervals. It also appears that the Kalahari linear dunes have been close to their threshold for reactivation throughout much of the late Quaternary, so that only minor changes in dune-forming agents, such as wind speed, plant cover, sand supply and effective precipitation, would have been needed to trigger renewed dune movement.

8.15 Desert dunes of Australia

The Australian sand deserts are the second largest sand deserts on earth and have been investigated in some detail, as will be evident from the following summary. Dunes and sand plains presently cover about two-fifths of Australia and have long attracted

attention. Until quite recently it was not known when dunes first appeared in Australia and when and under what conditions they had been active since then. Pollen evidence shows that the onset of climatic desiccation in central Australia dates back to the mid-Miocene some 15 million years ago (15 Ma) (Martin, 2006). The pollen data are entirely consistent with the evidence from dated molecular phylogenies of diverse Australian taxa indicating the greatest divergence of arid-adapted taxa around 15 Ma ago (Byrne et al., 2008).

The oldest stony deserts (known in Australia as ‘gibber plains’) have yielded cosmogenic nuclide ages of 4–2 Ma, possibly reflecting the onset of major Northern Hemisphere cooling and ice cap development over North America around 2.6 Ma (Fujioka et al., 2005). The earliest dunes did not appear until about a million years ago (Fujioka et al., 2009), reflecting an accentuation of the trend towards extreme aridity that was underway during the very late Pliocene and early Pleistocene (Fujioka and Chappell, 2010).

In the western Murray Basin of South Australia, Lomax et al. (2011) obtained 98 OSL ages from thirteen dune sections dating back to at least 380 ka. The ages showed two major phases of dune sand deposition at 72–63 ka and 38–18 ka, with some accretion at 14.5–13.5 ka, 12–11 ka and 8–5 ka. They concluded that although the dune records were discontinuous and often hard to interpret in terms of climate, high rates of dune deposition in this region tended to coincide with drier conditions, and breaks in the dune depositional record with wetter conditions inferred from other sources of evidence.

The distribution and orientation of the Australian desert dunes (Figure 8.14) have been mapped with increasing accuracy over the past few decades using a combination of ground surveys and air photos in earlier years (King, 1960; Jennings, 1968; Sprigg, 1979; Wasson et al., 1988) and satellite imagery and large scale topographic maps more recently (Hesse, 2010). Hesse’s (2010) remapping of desert dunes throughout Australia has convincingly demonstrated that there is a strong topographic control over the distribution of dune fields in Australia, with dunes preferentially occupying major depositional centres, much as in the Sahara. For example, the Lake Eyre Basin, which covers about one-seventh of the continent, is host to three major sand deserts: the Simpson, Tirari and Strzelecki deserts.

The very striking linear dunes of the Simpson Desert have been the subject of considerable fieldwork, and there has been a great deal of speculation as to how and when they developed. Several models have been proposed to account for linear dune formation in this region. Twidale (1972) concluded that the linear dunes of the Simpson Desert originated from transverse sand mounds or lunettes located on the downwind margin of widespread alluvial plains and playas. These source-bordering dunes then progressed downwind as linear dunes aligned parallel to the dominant sand-moving winds. He postulated that the dunes moved forwards by a process of downwind sand accretion and that they were able to move across many hundreds of kilometres of desert, given an adequate supply of sand. He observed that in many places the dunes

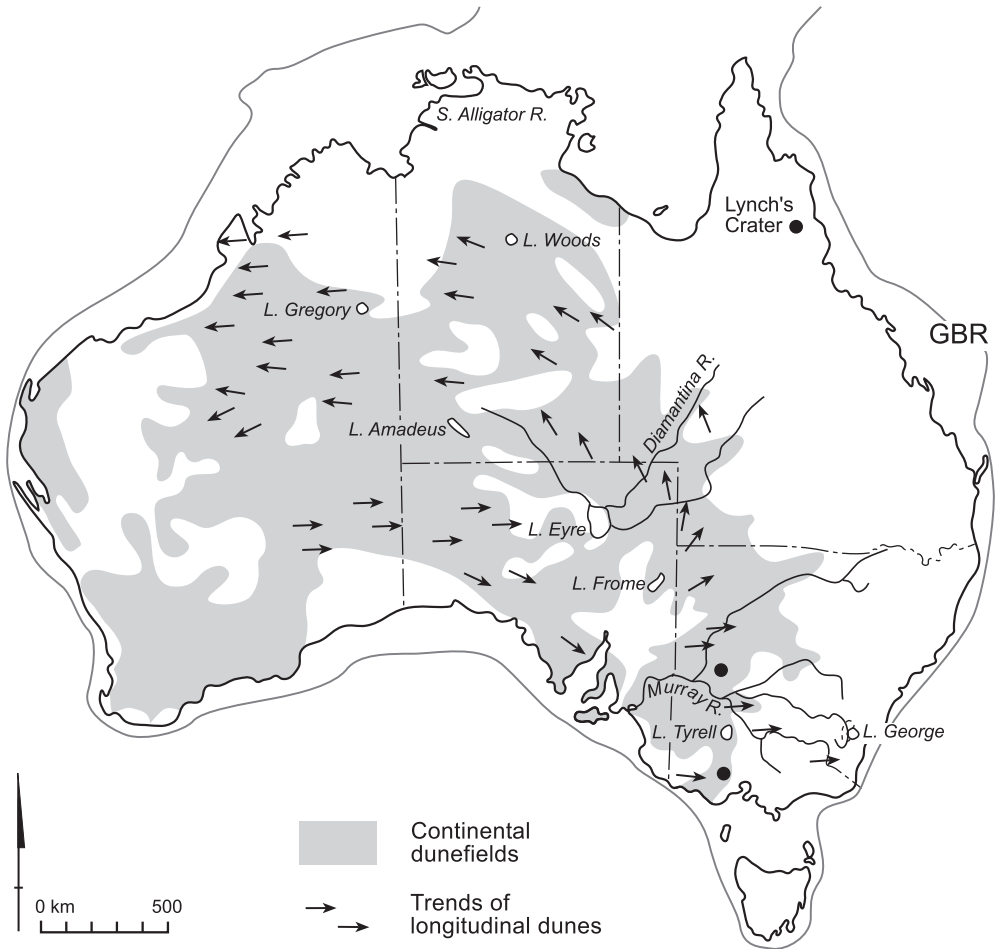


Figure 8.14. Map showing desert dunes, dune fields, lakes and rivers in the Australian arid zone.

were underlain by late Pleistocene alluvial and lacustrine sediments and concluded that, at least in south-west Queensland and South Australia, the linear dunes of the Simpson Desert were very late Pleistocene or early Holocene in age, although the chronological control at that time was rudimentary. Twidale later rectified this, and his subsequent work with Lomax et al. (2003) showed that dunes in the Strzelecki Desert yielded OSL ages of at least 65 ka, with a fluvial depositional phase of 160 ka, suggesting that far earlier phases of dune activity could not be ruled out.

Twidale's (1972) earlier conclusions in regard to long-distance eolian transport of sand have been somewhat modified by later work. Using a combination of physical and chemical analyses, including quartz oxygen isotope composition, Pell et al. (1999; 2000) concluded that the sands of the Simpson, Strzelecki, Tirari and Great Victoria

deserts were derived primarily from local bedrock with very little subsequent eolian transport. In those instances where the sand source was several hundred kilometres distant from the present-day dune field, transport had been dominantly fluvial and not eolian. They concluded that there had been some recent localised sediment input from modern fluvial systems in the Simpson, Strzelecki and Tirari deserts, while in Great Victoria Desert, Quaternary eolian transport or reworking of sand had been minimal.

The relationship between dune orientation and present-day wind direction is still unclear (Brookfield, 1970; Sprigg, 1979; Nanson et al., 1995), with most observers noting the broad general accordance between linear dune alignment and the general direction of wind flow associated with the anticyclonic air flow patterns over central Australia. Hollands et al. (2006) studied and dated linear dunes in the north-west Simpson Desert and concluded that there had been an approximately 160 km, or 1.5°, southward displacement of the sand-transporting wind system since the LGM. Sprigg (1979) had earlier argued that the dunes were active during glacial maxima and thus reflected the wind systems of those times. He believed that there had been an equatorward shift of more than five degrees of latitude in the southerly stream of dune-forming winds during the last glacial relative to the present interglacial in response to an intensification and northward displacement of the westerlies along the southern margin of Australia. This inference has received some support from the wind-blown dust record in marine cores off the east coast of Australia, with the marine record showing a threefold increase in dust flux during the LGM relative to the Holocene in temperate and tropical Australia (Hesse, 1994). We discuss the Australian desert dust record in [Chapter 9](#).

During and immediately after the LGM, dunes were active across the continent, including as far south as north-east Tasmania (Bowden, 1983; Duller and Augustinus, 1997; Duller and Augustinus, 2006). OSL ages for linear dunes in the Strzelecki and Tirari deserts are concentrated at 73–66, 35–32, 22–18 (LGM) and 14–10 ka (Rhodes et al., 2004; Fitzsimmons et al., 2007a). The two intervals with the most samples are at 20 ka, when sea level was 120 m lower than today and the Australian land-mass was about 25 per cent greater in area, and 14–10 ka, when temperatures were becoming warmer (Fitzsimmons et al., 2007a). The apparent gap between 20 ka and 14 ka suggests that few deposits have been preserved from that period. There was either little dune building at that time or subsequent reworking has removed the evidence. It will be interesting to see whether future closer sampling intervals for OSL dating of the desert sand dunes in Australia confirm or refute the various age clusters claimed by different workers.

Fitzsimmons et al. (2007b) obtained OSL ages for both transverse and linear dunes located on the eastern (downwind) margin of ephemeral Lake Frome immediately west of the arid Flinders Ranges. The ages for both types of dunes clustered at 66–57 and 22–11 ka. Transverse dune building began around 111–106 ka, while linear dunes

began to form at least 66 ka ago, with renewed activity at 43–28 ka, followed by soil development under more humid conditions. Concentrations of clay pellets within several horizons of both transverse and linear dunes reflected salt influx and episodic deflation from Lake Frome, prompting these workers to conclude that local hydrology rather than aridity had controlled dune initiation. Dune reactivation during the LGM began at around 22 ka and was associated with cold and regionally dry conditions and reduced plant cover.

Maroulis et al. (2007) and Cohen et al. (2010a) have studied the development of source-bordering dunes associated with changes in the fluvial regime of Cooper Creek in the Lake Eyre Basin of central Australia. Cooper Creek arises from the confluence of the Thompson and Barcoo rivers, both of which originate in the well-watered Eastern Highlands of Australia, where they depend on tropical summer rainfall. The Cooper then flows towards Lake Eyre in the heart of the arid zone, by which time it has lost most of its former discharge. Maroulis et al. (2007) developed a luminescence chronology for phases of vigorous fluvial activity and alluvial sand transport at intervals during Marine Isotope Stages (MIS) 8 to 3, with a long-term trend of progressively declining discharge during that time. Times of peak source-bordering dune activity were dated to late MIS 5 (85–80 ka) and mid-MIS 3 (50–40 ka), after which the flood plains became mantled with mud, leaving dunes as isolated features flanked by alluvial mud.

Nanson et al. (2008) conducted a comprehensive study of the middle and late Quaternary alluvial deposits in the lower 500 km of Cooper Creek, supported by eighty-five luminescence ages, both TL and OSL (see [Chapter 6](#)). They also found multiple episodes of enhanced flow during the last quarter million years and estimated that around 250–230 ka (MIS 7–6), mean bankfull discharge on Cooper Creek upstream of the Innamincka Dome was five to seven times larger than it is today, becoming less during and after the last interglacial (125 ka: MIS 5e). Strong flows continued to provide abundant seasonal sediment and to feed source-bordering dunes until about 40–35 ka (MIS 3), when the trend towards aridity became stronger.

Cohen et al. (2010a) concentrated their stratigraphic work on the low-gradient alluvial fan formed by Cooper Creek at the outlet of a narrow bedrock channel entrenched in the tectonically rising Innamincka Dome. They found that the base of the dune complex in this area dated back to at least 250 ka (MIS 7) and had been reworked in part by wind with sand replenishment from the river and associated source-bordering dune formation from around 120–100 ka, 85–80 ka and 65–53 ka. Contrary to expectation, the LGM was a time of enhanced river flow and sediment supply to the dunes from 28 to 18 ka. The evidence from boreholes and luminescence dating reveals a long history of river sands feeding transverse dunes that in turn gave rise to linear dunes by a process of vertical accretion, with negligible long-distance eolian sand transport.

8.16 Conclusion

Not a great deal of climatic information has yet come to light from studies of dunes alone. Studies of the dunes need to be buttressed by studies of intercalated fossil soils and of any associated fluvial and lake sediments and their associated fossils, as was carried out in exemplary fashion more than thirty years ago in the Nebraska Sand Hills (Ahlbrandt and Fryberger, 1980; Bradbury, 1980; Hanley, 1980). Although long-distance sand transport certainly seems possible given an adequate supply of sand, strong persistent winds and a lack of topographic obstacles, the sand deserts in the Sahara, Australia and China appear to be quite local features with relatively little long-distance transport of sand by wind. Desert dunes in general will not provide a reasonably continuous or long-term record of past climate. For this, we need to consult the record of wind-blown dust preserved both on land and in deep-sea cores, discussed in [Chapter 9](#), as well as the evidence from rivers and lakes.

9

Desert dust

The Lord shall make the rain of thy land powder and dust.
Deuteronomy 28.24

9.1 Introduction

Scientific interest in wind-blown dust has a respectable pedigree. In the late eighteenth century, Dr Matthew Dobson (1781) had already described the dust-transporting role of the Harmattan wind in West Africa, a topic revisited two centuries later by McTainsh (1980) and McTainsh and Walker (1982). In January 1832, Charles Darwin collected a sample of wind-blown dust using a gauze filter placed at the masthead of HMS Beagle while that vessel was anchored at Porto Praya in the Cape Verde archipelago off the west coast of the Sahara (Darwin, 1860, pp. 6–7). He sent this sample (and four other dust samples collected for him by the geologist Charles Lyell from a vessel several hundred kilometres further north) to the eminent German naturalist Professor Christian Gottfried Ehrenberg in Berlin. Ehrenberg identified no fewer than sixty-seven species of diatoms (Darwin’s ‘infusoria’) in the five samples, two of them marine and the rest freshwater (Darwin, 1846; Ehrenberg, 1851).

Darwin correctly attributed the dust to transport by the Harmattan wind that in winter blows from the Chad Basin across northern Nigeria and out across the Atlantic. The siliceous diatom frustules observed by Darwin and by Ehrenberg arise from the deflation of Holocene and older lake deposits in and around the Bodélé Depression in the Chad Basin (McTainsh, 1987; Washington et al., 2006). The African origin of wind-blown dust observed by vessels sailing across the Atlantic was thus recognised two centuries ago and has been the focus of attention ever since (Morales, 1979; Schütz et al., 1981; Williams and Balling, 1996, pp. 43–47; Goudie and Middleton, 2001; Prospero and Lamb, 2003; Goudie and Middleton, 2006; Goudie, 2008). This dust reaches as far as the Amazon, where it is an important source of plant nutrients

(Swap et al., 1992) and plays a significant role in the formation of ice crystals in clouds above the rainforest (Prenni et al., 2009).

It took somewhat longer for nineteenth-century geologists to appreciate that the loess deposits of Europe and North America were of eolian and not alluvial origin, as had been argued by Lyell (Pye, 1987, p. 198). In China, Ferdinand von Richthofen (1882) interpreted the loess deposits as eolian, a fact already well-appreciated by Chinese scholars 2,000 years earlier, as the late Professor Liu Tungsheng pointed out (Liu et al., 1985, p. 2). Richthofen's work eventually persuaded earth scientists that the loess deposits of Europe and North America were also eolian in origin.

In Australia, studies of wind-blown dust remained relatively neglected until the pioneering work by Butler (1956; 1974) on 'parna' (mantles of wind-blown silt and clay) in the Australian Riverine Plain and the recognition by Jessup (1960a; 1960b; 1961) that many of the soils in semi-arid South Australia were developed on parent materials formed primarily from wind-blown dust. Jessup's argument was qualitative but logical. He noted that the red-blown clays he had mapped over vast areas were clay-rich and uniform in colour and texture across the entire landscape, irrespective of underlying bedrock lithology or local relief. Later workers have built on the pioneering fieldwork of Butler and Jessup using a battery of techniques, including stable isotope analysis, trace element geochemistry and clay mineral analysis to identify the eolian dust component in soils and alluvial sediments (Chartres et al., 1988; Greene et al., 2001; Gatehouse et al., 2001; Mee et al., 2003). Somewhat similar arguments to those proposed by Jessup (1960a; 1960b; 1961) half a century ago, bolstered by detailed grain-size analysis, were advanced more recently in support of an eolian origin for the Pliocene Red Clay underlying the Quaternary loess in China (Lu et al., 2001).

Maher et al. (2010, p. 62) define dust in very general terms as 'wind-borne mineral aerosol'. Kukla (1987, p. 191), in reviewing the loess stratigraphy of central China, defined loess as 'a silt, transported and deposited by wind, loosely cemented by a fine syngenetic carbonate incrustation, formed in semi-arid continental climates'. Other workers have noted that close to source, the loess or dust can be relatively coarse, often with a modest amount of very fine sand particles as well as silt and clay carried in suspension by turbulent winds, but after several thousand kilometres of wind transport, much of the material still in suspension is dominantly fine silt and clay size, that is, finer than about 3.5 μm in diameter (Schütz et al., 1981; Coudé-Gaussen and Rognon, 1983; Liu, 1985; Liu, 1987; Liu, 1991; Goudie and Middleton, 2001). For our purposes, the definition of Maher et al. (2010) is perfectly adequate and avoids any prejudgement about possible origins.

A further question is: What constitutes a dust-storm? Early efforts focussed on whether or not a specified object located at a particular distance, for example, 1 km, ceased being visible during a dust-storm event (the *brume sèche* of French meteorologists: Bertrand et al., 1974; Bertrand, 1976). For the sake of simplicity, we can define a dust-storm as one in which the visibility is reduced to less than 1 km.

With the advent of the Total Ozone Mapping Spectrometer (TOMS), more consistent global measurements of dust plumes became possible, and it also became possible to follow dust plume movements over time (Goudie and Middleton, 2001; Maher et al., 2010). A dust plume shows the generalised path followed by dust-storms.

The aims of this chapter are to examine the origin, sources and distribution of eolian desert dust around the world and to evaluate the environments in which the dust is mobilised, transported and deposited. We then illustrate how this information has been used to reconstruct past environmental (and climatic) fluctuations with the evidence preserved in both terrestrial and marine sediments.

9.2 World distribution of desert dust plumes

Changes in the amount of wind-blown dust mobilised, transported and deposited on land and sea are both a cause and a consequence of global and regional climatic change (Harrison et al., 2001; Arimoto, 2001; Goudie and Middleton, 2001; Goudie and Middleton, 2006; Goudie, 2008; Maher et al., 2010; McGee et al., 2010). The present-day dust flux is variably estimated at between 1 and 3.5 Pg yr⁻¹, of which between 0.3 and 2 Pg yr⁻¹ are deposited over the ocean (1 Pg is 1×10^{15} g). A number of major dust plumes have been identified using satellite imagery, with two major source areas identified as the deserts of central Asia, north-west of the Loess Plateau of China (Liu, 1985; Kukla, 1987; Kohfeld and Harrison, 2001a; Pullen et al., 2011), and the Bodélé Depression in the Chad Basin of the southern Sahara (Goudie and Middleton, 2001; Goudie, 2008). Smaller plumes emanating from Patagonia, south-west Africa, and north-west and south-east Australia have also been recognised (Péwé, 1981; Kohfeld and Harrison, 2001a). Figure 9.1 shows the main dust source areas and the general directions of dust transport. However, this is to present a somewhat oversimplified picture of dust movement.

In the case of North Africa, for example, Coudé-Gaussen and Rognon (1983) have distinguished four main dust trajectories. One flows to the west along the southern margin of the Sahara, extending across the Atlantic via Barbados and across the equator to the Amazon Basin in South America. Another flows west across the northern Sahara and out into the Atlantic, eventually reaching North America. A third dust path flows northwards across the Mediterranean to reach southern France and the Alps. In exceptional cases, this plume may reach England (Pitty, 1968; Goudie and Middleton, 2006), Ireland and even Sweden (Franzén et al., 1994). A fourth North African dust path flows from the central and northern Sahara across Egypt and the Mediterranean to Israel and Arabia. Moreno et al. (2002) have examined the late Quaternary Saharan dust record preserved in a marine core in the Alboran Sea located in the western Mediterranean between Spain and Morocco. They found that enhanced northward transport of Saharan dust coincided with North Atlantic Heinrich events (see Chapter 3) and with times of strengthened high-northern-latitude atmospheric circulation.

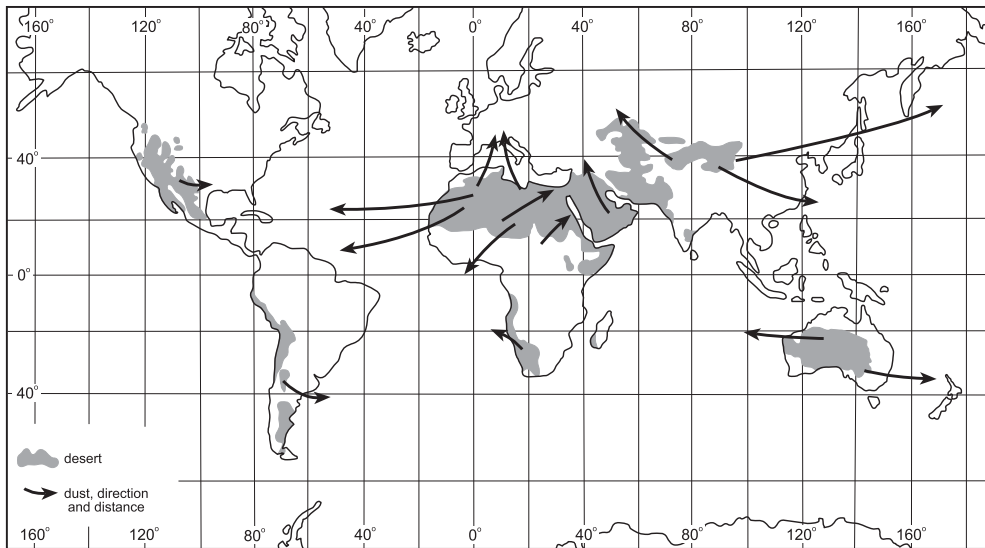


Figure 9.1. Map showing modern dust source regions and major directions of dust transport. (Adapted from Péwé, 1981, fig.1, and Harrison et al., 2001, fig.1a.)

In fact, during glacial stages, dust fluxes measured mainly in deep-sea cores were between three and five times greater than during interglacial times (Maher et al., 2010; McGee et al., 2010). In Antarctic ice cores, times of maximum dust accumulation were also synchronous with times of minimum temperature and with times of minimum carbon dioxide concentrations measured from trapped air bubbles in the ice (Petit et al., 1981; Petit et al., 1999; Jouzel et al., 2007). Disentangling a precise climatic signal from these fluctuations in dust accumulation rates is far from easy, given that dust production, transport and deposition are determined by a variety of factors, including changes in wind velocity and gustiness, source area and extent, vegetation cover, dust transport paths and deflation from glacial outwash deposits (Péwé, 1981; McTainsh, 1987; Maher et al., 2010; McGee et al., 2010).

9.3 Origins and physical characteristics of desert dust

There have been many attempts to distinguish between wind-blown dust particles formed as a result of desert weathering processes, such as salt-weathering or abrasion between moving sand grains, and those formed as a result of frost-shattering or glacial abrasion in glaciated and periglacial landscapes (Smalley and Vita-Finzi, 1968; Vita-Finzi and Smalley, 1970; Goudie et al., 1979; Pye, 1987). Wind is certainly competent to undercut small sandstone hillocks in the Sahara (Figure 9.2). Such efforts, while useful to our understanding of particle micromorphology (Coudé-Gaussen and Rognon, 1983), are somewhat chimerical when it comes to disentangling the origin of desert dust. Many deserts are flanked by high, glaciated mountain ranges, such



Figure 9.2. Isolated sandstone hillock undercut by wind erosion near In Guezzam, central Sahara. (Photo: J.D. Clark.)

as the Tian Shan north of the Tarim Basin in western China, the Andes west of the Patagonian Steppes, the Sierra Nevada and the Rockies flanking the deserts of the south-western United States and the Atlas Mountains north-west of the Sahara. Others are dotted with uplands that were themselves prone to glacial and periglacial activity, such as the Hoggar Mountains in the north-central Sahara.

There is no good reason why angular silt-sized particles cannot be produced within deserts by a variety of physical and chemical weathering processes, including glacial grinding and frost shattering in desert mountains, salt weathering (Goudie et al., 1979), chemical weathering of the bedrock (McTainsh, 1987) and abrasion during turbulent sand storms (Amit et al., 2009). In addition, as Darwin observed correctly in 1832, some of the silt-sized particles could be the siliceous frustules of diatoms that lived in the early to mid-Holocene freshwater lakes that once flourished across the Sahara. Furthermore, there is no particular reason why the dust deposited in any one locality cannot have come from several different sources. Another complicating factor is the remobilisation of dust mantles laid down along the path of the dust plume (McTainsh, 1987).

In the Negev Desert of Israel, the loess often consists of a relatively coarse quartz fraction (50–60 μm) with OSL ages dating back to about 180 ka, in a matrix of much finer particles (3–8 μm) (Crouvi et al., 2008; Crouvi et al., 2009). The finer material is

considered to have blown in from the Sahara and so may have travelled thousands of kilometres in suspension in the atmosphere. In contrast, the coarser particles are now believed to be of more local origin, having formed as a result of sand-particle abrasion during the eastward progression of the dunes in the Sinai Desert (Crouvi et al., 2008; Crouvi et al., 2010; Enzel et al., 2010). The source of these latter dunes is the Nile, and the sands are thought to have blown inland from deltaic sediments exposed during times of lower glacial sea level (Amit et al., 2009). One unresolved issue with this interpretation is why the coarse-grained loess is, according to present evidence, dated by OSL as no older than about 180,000 years, particularly given that the Nile has apparently been transporting sediments to the delta for more than 2 million years (Williams and Talbot, 2009; Davis et al., 2012).

The dominant component of most desert dust plumes is quartz, with lesser quantities of other minerals, such as feldspar and variable amounts of goethite, haematite and other iron minerals. The desert dusts of central Asia are enriched in illite, but illite forms a very minor component of the dusts of the Near East and North Africa (Singer, 1988). Dust blowing off continental shelves exposed to deflation as a result of glacially lowered sea levels often contains relatively abundant particles of calcium carbonate, which can be transported for hundreds or even thousands of kilometres from source to be later incorporated into soils and sediments well inland or far-removed from the source (Williamson et al., 2004; Dart et al., 2007; Amit et al., 2010).

After initial fallout of the coarser grains, the modal diameter of particles carried long distances (i.e., in excess of 1,000 km) tends to remain fairly constant at 3.5–2.5 μm . Deposition of the coarse grains is relatively rapid and is governed by Stokes' Law:

$$W = 2/9 gr^2(p - p_a/M) \quad (9.1)$$

In this expression, W is the terminal velocity of a particle of radius, r , and density, p , falling through still air (density p_a and viscosity M), and g is the acceleration due to gravity. This relation is also true of volcanic dust plumes in which only very fine particles in the size range of 0.5 to 2 μm are likely to persist in the stratosphere for periods of several or more years (Lamb, 1972, p. 411).

9.4 Dust entrainment and transport

The detachment and entrainment of silt-sized dust particles is a function of wind velocity and wind gustiness, and so is tightly controlled by local and regional synoptic conditions. Among the other factors that control the potential availability of dust are aridity, soil type, plant cover and surface roughness. As with desert dunes, we are again faced with a multivariate problem, so that it is often hard to separate out which are the more influential factors responsible for dust entrainment. Dry soils or sediments and sparse or absent plant cover are obviously important prerequisites, but without strong

winds, they remain passive factors. Many authors have stressed the importance of high wind velocities in mobilising dust from the ground surface (Maher et al., 2010), but a more subtle interpretation involves wind gustiness (McGee et al., 2010).

McGee et al. (2010) noted that the glacial dust flux in the late Pleistocene was roughly two to four times that of the interglacial dust flux in many parts of the semi-arid world. They argued that enhanced glacial wind gustiness resulting from steeper pole to equator thermal gradients was the dominant, or first-order, driver of such an increase. They focussed their discussion on East Asia (north-western China), North Africa (the Chad Basin) and southern South America. Using a threefold approach (paleoenvironmental data; modern synoptic controls of dust-storms; global atmospheric models), they considered and discounted a number of other possible causes of enhanced dust flux, including changes in source area, sediment supply, plant cover, aridity, atmospheric carbon dioxide concentration ($p\text{CO}_2$) and sea level. However, they did acknowledge some local influences involving sediment supply and vegetation cover.

Modern dust-storms have thrown some light on this problem. Liu Tungsheng and his colleagues (1985, pp. 149–157) have analysed the meteorological conditions under which dust is mobilised from eastern Siberia and transported as far as eastern China, Korea and Japan. Strong frontal winds, high wind velocities and powerful convective updrafts are associated with initial mobilisation and dust entrainment, while jet stream activity seems implicated in long-distance transport.

A well-known modern dust-storm event in the Nile Valley is the *haboob*, of which there are three main types, each of which is associated with quite distinct synoptic conditions (Griffiths and Soliman, 1972, p. 93). Haboobs are associated with high wind velocities (55 km/hour and more) and substantial turbulence, vindicating the gustiness hypothesis, but their location is governed by a suitable supply of silt-sized particles on the ground (Kendrew, 1961, p. 71). Such dust-storms attain heights of at least 1,500 m and advance along a sharp front some 25 km wide at a rate of about 55 km/hour.

The Australian dust record is considered in detail in Section 9.7, so it will be enough to say here that peak dust flux coincided with the LGM in marine cores to the east (Hesse, 1994; Hesse et al., 2004) and south of Australia (Gingele and De Deckker, 2005), as well as in the lunettes of the Willandra Lakes of arid western New South Wales (Bowler, 1998). It also appears that the dust source areas have not remained constant over time. For example, Hesse and McTainsh (2003) found that the northern limit of the dust plume was 350 km, or 3°, north of the present limit during the interval from 22 ka to 18 ka ago, indicating a major expansion in source area.

Nor should the role of aridity in preparing areas for deflation be minimised. The Chad Basin was hyper-arid during the LGM and had a vastly expanded area available for deflation, for which there is very strong local and regional evidence (Servant, 1973; Servant and Servant-Vildary, 1980; Hoelzmann et al., 2004). It seems very

probable that enhanced LGM aridity in the Bodélé Depression and adjoining region within the Chad Basin would have significantly increased the potential supply of dust to the Harmattan winds of that time.

The present is also a useful guide to the past. Work over the past fifty years has shown that years of drought in the Sahel zone of West Africa are followed by an increase in dust flux from the southern Sahara across the Atlantic (McTainsh, 1985; Middleton, 1987; Pye, 1987; Williams and Balling, 1996; Goudie and Middleton, 2001; Prospero and Lamb, 2003). This work indicates that both aridity and the concomitant reduction in annual plant cover associated with drought conditions have a major influence on the volume of dust blown from the source area. The observations of McTainsh in northern Nigeria have indicated that although the maximum particle size of dust carried by the Harmattan winds diminishes slowly downwind, it can diminish very rapidly laterally, perpendicular to the long axis of dust transport (McTainsh, 1980; McTainsh and Walker, 1982; McTainsh, 1984). As noted in Chapter 8, there was also increased LGM aridity in the Lake Eyre Basin in Australia, another major source of dust in both the past and present.

It thus seems safe to conclude that aridity, suitable surface soils and a sparse plant cover are necessary preconditions for dust entrainment and that given such conditions, strong and gusty winds will initiate particle movement. Once in suspension, strong unidirectional winds will ensure transport of the dust particles well-beyond the source area.

9.5 Dust deposition and accumulation

Figure 9.3 shows the distribution of loess deposits across the globe. Dust in suspension in the atmosphere can fall back to earth as dry dust fall, operating under Stokes' Law, as described in Section 9.3. Alternatively, it can be removed from the air during rainfall events. Some of this dust may already have been present as nuclei for water or ice droplets in clouds (Arimoto, 2001; Prenni et al., 2009), but most of it is scavenged from the atmosphere during rainstorms that frequently follow on the heels of major dust-storms. Once the dust has reached the ground, it may be remobilised by wind gusts, as in northern Nigeria during the Harmattan season (McTainsh, 1987) or, if suitable dust traps are available, it will remain on the surface and ultimately form a dust mantle or loess deposit. Many authors have stressed the importance of dust traps for desert loess formation (Coudé-Gaussen and Rognon, 1983; Tsoar and Pye, 1987; Coudé-Gaussen et al., 1987; Williamson et al., 2004). In fact, Coudé-Gaussen and Rognon (1983) have argued that a 'pluvial' climate and a dense cover of grasses and shrubs were necessary for loess accumulation in the Matmata limestone uplands of southern Tunisia during the late Pleistocene. In the Loess Plateau of north-central China, the reverse pattern seems to have prevailed, with maximum rates of loess accumulation during the colder, drier, windier glacial phases of the Quaternary and

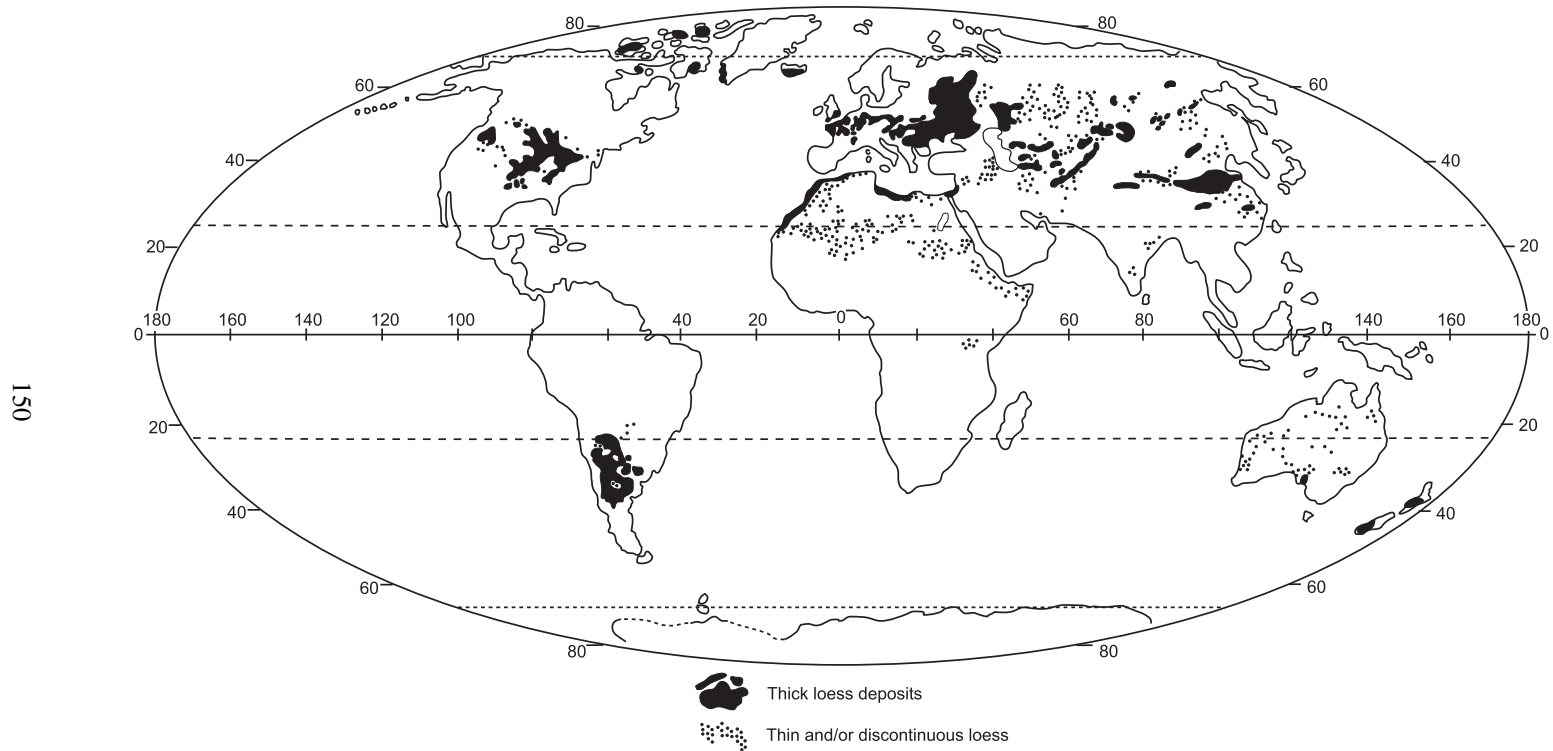


Figure 9.3. Map showing the global distribution of loess and desert dust. (After Williams et al., [1998](#).)

reduced rates of accumulation and widespread soil development during the warmer wetter interglacial phases (Liu, 1985; Liu, 1987; Liu, 1991).

Not all loess mantles are in primary context, and once deposited on hill slopes and valley sides, the loess will be subject to erosion and movement downslope under the influence of mass movement and running water, as discussed in [Chapter 10](#). Indeed, many of the fine-grained, late Pleistocene valley-fill deposits in the mountains of the Namib Desert (Eitel et al., 2001; Heine and Heine, 2002; Eitel et al., 2005), the Sinai (Rögner et al., 2004), the Negev (Avni, 2005; Avni et al., 2006), the presently semi-arid Flinders Ranges of South Australia (Williams et al., 2001; Chor et al., 2003; Williams and Nitschke, 2005; Williams and Adamson, 2008; Haberlah et al., 2010a; Haberlah et al., 2010b), as well as the Matmata Hills of Tunisia (Coudé-Gaussen et al., 1987) and the semi-arid Chifeng region of Inner Mongolia in northern China (Avni et al., 2010) consist primarily of reworked loess, and so have more to tell us about fluvial activity at that time than about causes of primary dust deposition.

These fine-grained, late Pleistocene valley fills are up to 20 m thick and are widespread within dissected arid uplands in Africa and Australia. Such deposits are not accumulating today (Williams et al., 2001). The dominant lithology is silty clay, with minor lenses of fine to medium gravel. In the arid Flinders Ranges of South Australia, the valley fills consist of a massive lower unit and a finely laminated upper unit. The lower unit contains abundant unbroken ostracod and aquatic mollusc shells, indicating deposition under perennially wet conditions. The uniform stable carbon and oxygen isotopic composition of the shells also confirms deposition under stable climatic conditions (Glasby et al., 2007). In contrast, the individual fining-upwards laminae in the upper unit contain broken shells and fragmented plant remains, and are best interpreted as slackwater deposits (Haberlah et al., 2010a; Haberlah et al., 2010b). The valley fills have been dated using paired charcoal and shell samples for AMS ^{14}C analysis combined with optically stimulated luminescence dating. They were laid down between about 35 ka and 15 ka (Williams et al., 2001; Glasby et al., 2010; Haberlah et al., 2010a; Haberlah et al., 2010b).

Particle size analysis, strontium isotope analysis and rare earth element composition show that the silty clays in the Flinders Ranges valley fills were largely derived from reworked loess. The loess was blown in from the south-west by stronger westerlies and accumulated along the ridges and slopes of the north-south aligned ranges. The atmospheric carbon dioxide concentration at this time was as low as 180–200 ppmv. Such low concentrations favoured the expansion of grasses and herbs at the expense of eucalypt trees (Williams and Adamson, 2008). The grass cover would have provided an effective dust trap. The demise of the deep-rooted river red gums (*Eucalyptus camaldulensis*) along the valley bottoms would have resulted in a slow rise of the local water-table, leading to swampy conditions. Low summer temperatures and much reduced evaporation were conducive to persistence of these wetlands (Chor et al., 2003). A much weaker summer monsoon regime in the tropical north would have

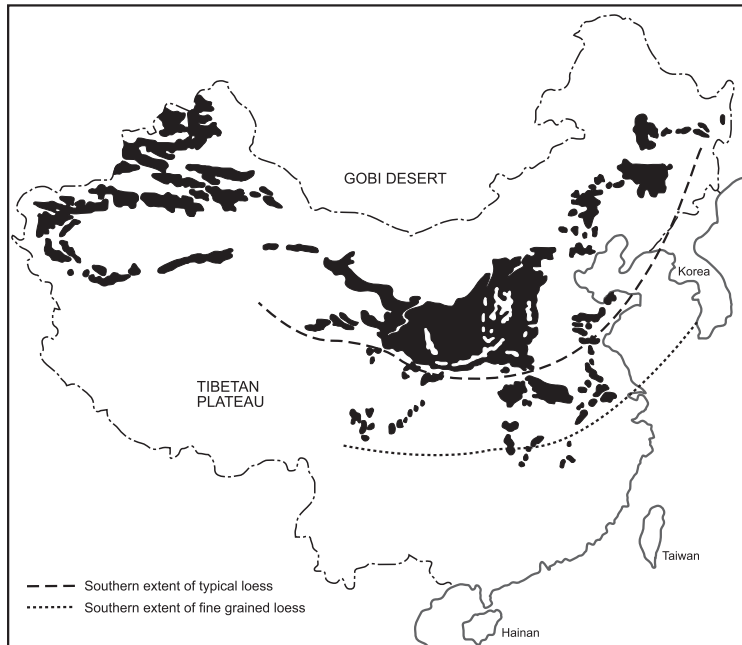


Figure 9.4. Map showing the distribution of loess in China. (After Williams et al., 1998.)

meant fewer incursions of high-intensity, potentially erosive summer rain. By analogy with the present, the winter rains were most likely prolonged and gentle, enabling the thin loess deposits to be washed off the slopes and down to the valley bottoms by overland flow (Williams et al., 2001).

A major environmental change took place close to the onset of the Last Glacial Maximum (21 ± 2 ka), after which episodic floods deposited the laminated unit (Haberlah et al., 2010a; Haberlah et al., 2010b). These findings show that the LGM climate was not uniformly cold and arid, and in fact experienced significant decadal scale variability. Abrupt incision after 15 ka denoted a return to a high-intensity summer rainfall regime and the end of the wetlands (Williams et al., 2001; Williams et al., 2009b).

9.6 The Loess Plateau of China

By far the best terrestrial record of wind-blown dust accumulation is that preserved in the Loess Plateau of China (Figure 9.4), which covers an area of about 440,000 km² and is mantled by loess that reaches up to 350 m in thickness but is usually only about 100 m (Liu, 1985; Liu, 1987; Liu, 1991; Kukla, 1987). Wind-blown dust began to accumulate in China early in the late Miocene, some 8 million years ago (8 Ma) (An and Porter, 1997) and possibly even as early as 22–24 Ma ago (Guo et al., 2002; Sun et al., 2010).

The brilliant work by Professor Liu Tungsheng and his colleagues demonstrated that beds of unweathered loess laid down during cold, dry climatic episodes alternated with beds of highly weathered loess in which often well-developed soils (Kemp, 2001) had formed during warm, wet climatic interludes, as indicated by the pollen and fossil faunal remains (especially mollusca) preserved within them (Kukla, 1987; Liu and Ding, 1998; Kohfeld and Harrison, 2001b). A simple notation system was developed in which L refers to fresh loess beds and S to fossil soils, numbered from the Holocene (S1) back through the LGM (L1) to the base of the sequence. Magnetic susceptibility analyses revealed that the red paleosols contained iron formed during pedogenesis – iron lacking in the unweathered loess beds (Kukla, 1987; Evans and Heller, 2001; Maher et al., 2010). Particle size analysis showed that coarser grains were more abundant during the cold, dry phases of loess deposition. Source areas were the Taklamakan Desert and several sand deserts in the Gobi and north-west Inner Mongolia, including the Badain Jaran, Tengger and Mu Us deserts (Ding et al., 1999), as well as central Asia (Sun et al., 2010).

The initial interpretation of the alternating loess and soil (L/S) sequence was that the soils developed during times when the summer monsoons were at least as strong as they are today (i.e., during interglacials and interstadials), and the loess units developed during times when the winter monsoon was stronger and the Siberian High Pressure system was more intense than they are today (i.e., during glacials and stadials). Roe (2009) has challenged this interpretation, arguing that most of the present-day dust outbreaks from western China occur in spring and are associated with the passage of strong cold fronts that produce intense windstorms able to entrain substantial amounts of dust. The same is true in Mongolia, where the dust-storms occur primarily in the spring (Middleton, 1991). These events coincide with weakening of the Siberian High, and this pattern of events is also likely to have occurred in the past, suggesting that the simple dichotomy between ‘winter’ and ‘summer’ monsoon is somewhat oversimplified. Another problem relates to differential preservation of the loess sequence, with episodes of intense gully erosion coinciding with the onset of interglacial phases (Porter and An, 2005).

Detailed sampling of stratigraphic sections located on a set of east-west and north-south transects has revealed a sequence of thirty-seven loess-soil couplets spanning the past 2.5 Ma. Each couplet represents a cold and dry phase of rapid dust accumulation and an ensuing wet and warm phase of weathering and soil formation. A soil is defined as weathered loess showing at least as much pedological organisation as the widespread early Holocene soil at the top of the loess sequence. Interpretation of the loess-soil couplets is based on high-resolution sampling and detailed analyses of grain size, magnetic susceptibility, organic carbon, sediment micromorphology and mineralogy, calcium carbonate content and mollusc species. At a very general level, the soils are thought to indicate a weaker winter monsoon and a stronger summer monsoon. Conversely, the coarser-grained, unweathered loess with generally much

weaker magnetic susceptibility values is regarded as evidence of a stronger and more extensive winter monsoon and a weaker summer monsoon. Chronological control is based on the paleomagnetic time scale, cross-correlation with the marine oxygen isotope record and a combination of radiocarbon and thermoluminescence dates for the more recent part of the sequence.

Further refinements to our understanding of the unrivalled Chinese loess sequence include detailed attention to dust source areas (Ding et al., 1999; Pullen et al., 2011) and to the topographic controls over dust deposition (Sun, 2002), as well as a finer resolution OSL chronology of key loess sections (Lu et al., 2007). There have been promising attempts to correlate the loess record with climatic events recorded in Greenland ice cores and North Atlantic marine cores (Porter and An, 1995), as well as with temperature changes inferred from the oxygen isotopic record obtained from foraminifera in marine cores collected from the East China Sea (Liu et al., 1985; Kukla, 1987) and the North Pacific (Hovan et al., 1989). At all events, it seems that times of maximum loess accumulation in China coincide with times of greater aridity, expanded source areas, reduced plant cover and both stronger and gustier wind regimes.

9.7 Wind-blown dust in Eurasia, Africa, America and Australia

The well-studied loess deposits of America are not always strictly desert dust deposits, given that they are often associated with deflation from the outwash plains associated with meltwater run-off from the great Laurentide and Cordilleran ice caps, which reached their most recent maximum extent during the LGM. However, in presently semi-arid Nebraska, rates of loess mass accumulation obtained from luminescence dating were exceptionally high between 18 and 14 ka (Roberts et al., 2003). These authors further suggested that the high atmospheric dust loading over that area implied by these rates may have influenced radiative forcing sufficiently to have contributed to the several thousands of years of colder-than-present climate over central North America at that time, despite higher-than-present summer insolation values. Much as in China, the loess deposits of Russia and central Asia consist of alternating loess units and buried soils, some of which form composite and other discrete soil layers (Rutter et al., 2003). The loess accumulated during colder, drier, windier episodes when the desert source areas were greatly expanded and frost action was pronounced across the landscape. Patagonia was a major dust source for Antarctica during glacial times, when the source area had expanded twofold as a result of lower glacial sea levels. Once meltwater lakes had formed on the Patagonian outwash plains at the foot of the glaciated mountains, the supply of wind-blown dust was rapidly curtailed (Sugden et al., 2009; Maher et al., 2010).

In Australia, the two main source areas for wind-blown dust were (and are) the Lake Eyre Basin and the Murray-Darling Basin (McGowan et al., 2005; Petherick

et al., 2009). Hesse and McTainsh (2003) have reviewed work on modern dust-storms and Quaternary eolian dust in marine cores. The marine record shows that the rate of dust accumulation during the LGM was three times greater than it was during the Holocene (Hesse, 1994). Hesse and McTainsh (2003) attributed this to weakened Australian monsoon rains in the tropical north and a drier westerly circulation in the temperate south. From 33 ka to 16 ka, there was a strong flow of dust to the east and south over the southern half of the Australian continent. The northern limit of the dust plume in eastern Australia appears to have extended about 350 km, or 3°, north of the present limit from 22 to 18 ka (Hesse, 1994; Hesse et al., 2004). This is consistent with a northward shift of the high-pressure subtropical ridge (STR) in glacial times to its present summer location near 35°S. The STR separates the tropical easterly circulation from the mid-latitude westerlies (Hesse, 1994; Hesse et al., 2004).

Further evidence of major dust deflation during the LGM comes from Pleistocene Lake Mungo in semi-arid western New South Wales. With a chronology solidly founded on more than 200 ¹⁴C, TL and OSL ages, Bowler (1998) and Bowler and Price (1998) showed that eolian dust (*Wüstenquartz*, or desert quartz dust) began to accumulate in the lunettes on the eastern side of Pleistocene Lake Mungo and adjacent lakes from about 35 to 16 ka, with a peak centred around the LGM, when clay dunes and gypseous lunettes were actively forming on the downwind margins of seasonally fluctuating lakes in many parts of semi-arid south-east and south-west Australia immediately before and between 21 and 19 ka (Williams et al., 2009b). Major deflation of exposed lake-floor sediments coincided broadly with the time of extreme aridity centred on the LGM (e.g., Lake Eyre: Magee and Miller, 1998). In two marine cores off the coast of South Australia, Gingele and De Deckker (2005) found evidence of enhanced wind-blown dust deposition at roughly 70–74 ka, 45 ka and 20 ka, all times of minimum insolation in these latitudes. These were also periods of widespread lake desiccation, dune building and sparse vegetation cover in central and southern Australia (Croke et al., 1996). Fine resolution analysis of a late Quaternary dust record from eastern Australia offers support for the existence of two stadials during the LGM (Petherick et al., 2008).

9.8 Influence of desert dust on local and regional climate

Except for the deserts of the North American Southwest (see Chapter 20), glacial maxima were in general drier than today and interglacial maxima were as wet or wetter. The deserts of North Africa, Arabia, central Asia, China, Patagonia and Australia all display evidence of more vigorous eolian dust flux during glacial maxima. Nor should it be forgotten that eolian dust might have an influence on local and regional climates. In tropical West Africa, from 15 ka to about 7 ka, the rivers were mainly depositing clays, and after 7 ka they mostly carried sands. Maley (1982) attributed this abrupt

hydrologic change to a change in the size of raindrops. Abundant atmospheric dust would provide the nuclei for many small raindrops to form, resulting in gentle, non-erosive rains. Conversely, a reduction in atmospheric dust load would result in large, highly erosive drops. The momentum of a falling raindrop is the product of its mass and velocity, so this interpretation is physically plausible.

Other workers have noted that the very high inputs of eolian dust to central Antarctica and Greenland during the Last Glacial Maximum, which are clearly evident in the Antarctic (Petit et al., 1981; Petit et al., 1990; EPICA Community Members, 2004; Jouzel et al., 2007) and Greenland ice cores (Svensson et al., 2000; Ruth, 2005), are consistent with shorter dust wash out times and a weaker global hydrological cycle. It is also possible that a high concentration of atmospheric dust may in itself have contributed to the lowering of sea surface temperatures evident in the tropical western Pacific, especially in the warm shallow seas, or *West Pacific Warm Pool*, immediately to the north of Australia. Given the growing recognition of the interactions between present-day desertification processes, dust generation and the impact of dust particles on scattering incoming solar radiation, it seems highly plausible that wind-blown dust would be both a cause and an effect of Quaternary climatic fluctuations (McTainsh, 1989; Harrison et al., 2001; Kohfeld and Harrison, 2001b; McTainsh and Lynch, 1996; Maher et al., 2010). An interesting illustration of this is the 200-year record of wind-blown dust immediately following the 74 ka Toba volcanic super-eruption in Sumatra identified by Zielinski et al. (1996) in the Greenland GISP2 ice core. Because none of the preceding or following stadials showed a comparable dust signal, it is probable that volcanic cooling and drought triggered by this eruption may have had a global impact (Williams et al., 2009a). At the very least, there was sufficient climatic impact from this eruption for dust to be mobilised from hitherto stable soil surfaces in central Asia.

Interpreting the dust record from marine sediment cores is not always straightforward (Stuut and Lamy, 2004). Certain desert rivers flowing into the ocean may be carrying sediments eroded from fine-grained valley-fill deposits that were themselves originally derived from reworked loess, as in the Namib Desert of south-west Africa. It is therefore important to avoid reliance on the putative dust record alone where contamination from other sediment sources might have occurred (Gasse et al., 2008). However, where the only plausible sediment source is likely to have come from continental eolian dust, as off the coast of Mauritania, variations in the dust flux may provide useful first-order climatic information (deMenocal et al., 2000). In this study, a sharp decline in the dust flux after 14.8 ka coincided with the abrupt return of the summer monsoon in tropical Africa and the onset of the so-called 'African Humid Period' (deMenocal et al., 2000). The increase in dust flux from 5.5 ka onwards coincided with desiccation over the southern Sahara, southward displacement of the Intertropical Convergence Zone and weakening of the summer monsoon.

9.9 Conclusion

Fine-grained, wind-blown dust accumulated at intervals throughout the Quaternary on the downwind margins of deserts in Africa, Australia, South America and Asia (Pye, 1987). In terms of their sorting and mineral composition, they are virtually identical to the central European and North American loess mantles, which accumulated downwind of the fluvio-glacial outwash plains, so the term loess is used here for any fine-grained eolian deposit, irrespective of its original provenance. The eolian dust deposits in the Loess Plateau of central China are the thickest and most extensive loess deposits in the world (Liu, 1985; Liu, 1987; Liu, 1991; Kukla, 1987). They cover an area of 440,000 km² and attain thicknesses commonly in excess of 100 m and more than 300 m near the city of Lanzhou. Detailed studies over the past three decades have made the Chinese loess sequence with its alternation of unweathered loess and intercalated soils one of the most informative continental sequences covering the last 2.5 Ma that exists on earth (Liu, 1991). Comparison of the Chinese loess record with evidence from deep sea cores and the Greenland and Antarctic ice cores strongly confirms the climatic interpretation of the loess-soil couplets, with glacial maxima synchronous with times of maximum dust deposition and interglacials synchronous with times of maximum weathering and soil development. The loess sequence in China illustrates how accurate dating and careful evaluation of different lines of evidence is essential in reconstructing environmental change in deserts.

Finally, it is worth noting that although desert dust may prove a boon to those of us seeking to reconstruct past climatic changes in deserts, it is a persistent bane to the present-day inhabitants. I leave the last words to Kendrew (1957, p. 215): ‘Dust, not rain, is the great discomfort of life in arid lands. Except on still nights the air is full of fine particles which percolate through the finest chinks into houses and even closed boxes. Dust lies thick on every shelf, covers furniture, settles on food, and is inhaled in the air we breathe’. John Steinbeck makes a very similar point in *The Grapes of Wrath* (1939), discussed in Chapter 24.

10

Desert rivers

Hence in regions of small rainfall, surface degradation is usually limited by the slow rate of disintegration; while in regions of great rainfall it is limited by the rate of transportation.

G.K. Gilbert (1843–1918)

Report on the Geology of the Henry Mountains (1877, p. 105)

10.1 Introduction

During the first seven months of 1999, the arid Alashan Plateau of Inner Mongolia in northern China was still in the grip of one of the worst droughts in human memory. The entire region had endured eight years of below-average precipitation, and the local Mongolian herdsmen were becoming increasingly worried on behalf of their camels, sheep and goats. The drought effectively ended at midday on 5 August 1999, when a sudden thunderstorm in the mountains of the western Alashan unleashed more than 300 mm of rain over the next thirty hours (Williams, 2000a). This would be sufficient to replenish water levels in the piedmont wells and to sustain plant growth for at least three more years, according to the elderly Mongolian farmer and his wife in whose home we had taken refuge. We had been crossing the wide, boulder-strewn bed of an ephemeral stream channel which had barely flowed for several years but soon became a raging torrent carrying football-sized granite boulders along its bed.

At about the same time, more than half a world away, unusually heavy rains fell onto the parched clay plains of the central Sudan, flooding the hollows between the dunes immediately east of the White Nile, triggering sheet-floods from the low upland ridge separating the valleys of the lower Blue and White Nile rivers (Williams and Nottage, 2006) and causing breaching and overflow from the main canal feeding into the most enduring and successful large irrigation scheme in Africa – the Gezira Irrigation Scheme in central Sudan.

During the early 1940s, two young German geologists, Henno Martin and Hermann Korn, spent two-and-a-half years living in the remote gorges of the Namib Desert and witnessed several flash floods, one triggered by a mere fifteen minute downpour:

The whole thing [rainstorm in the distant mountains] had lasted perhaps a quarter of an hour, and an hour later we heard the roaring of water rushing down the main gorge. We ran quickly to see it as it swirled along, a tumultuous frothing bore of brown water over two metres high, flattening the tough bushes in its path, uprooting trees and tossing them into the air like matchsticks. It seemed almost incredible that a short downpour could produce such a volume of water. (Martin, 1983, p. 270)

Not all desert rivers are presently as ephemeral as the Alashan and Namib examples just mentioned. The Nile is a well-known example of a desert river that originates well-outside the desert, bringing sediments from its mountainous headwaters to be deposited along its narrow flood-plain in hyper-arid Egypt. The ever-perceptive Herodotus (ca. 485–425 BC) commented more than 2,500 years ago on the Nile silt in Egypt, noting that ‘the soil of Egypt does not resemble that of the neighbouring country of Arabia, or of Libya, or even of Syria . . . but is black and friable as one would expect of an alluvial soil formed of the silt brought down by the river from Ethiopia’ (translated by Aubrey de Sélincourt, 1960, p. 106).

In fact, many desert rivers flow from well-watered uplands adjoining the desert and remain perennial today, like the Awash River that flows from the highlands of Ethiopia into the hyper-arid Afar Desert, finally disappearing into Lake Abhe, which is now a vast and very shallow lake. The spectacular algal limestone pillars that rise several score metres above lake floor level bear witness to times of higher lake level (Fontes and Pouchan, 1975; Gasse, 1975; Gasse, 1976) when flow in the Awash was greater than it is today. The Pliocene and Pleistocene alluvium in the now arid Middle Awash Valley (Figure 10.1) contains a wealth of fossils, including the well-known hominid fossils of *Australopithecus afarensis* and cognate discoveries (see Chapter 17).

Likewise, the Cooper and Diamantina rivers that flow from the Eastern Highlands of Australia towards Lake Eyre are but a shadow of their Pleistocene ancestors, as shown by their deep and extensive distal alluvial sands (Nanson et al., 1992; Cohen et al., 2010a) and, indirectly, by the high lake strandlines abutting the present salt lake (Magee, 1998; Magee et al., 1995; Magee and Miller, 1998; Magee et al., 2004). Today Lake Eyre only receives water from its feeder channels during exceptionally wet years, most notably during extreme La Niña events (Allan, 1985; Kotwicki, 1986; Allan et al., 1996; Kotwicki and Allan, 1998), including in 2011 and 2012. Indeed, once extensive freshwater lakes fed by formerly active river systems are a feature of many deserts (Mabbutt, 1977, pp. 262–271; Cooke et al., 1993, pp. 202–219). The extensive linear network of salt lakes in western Australia (Van de Graaff et al., 1977) provides an enduring legacy of the rivers that flowed when Australia and Antarctica were part of a single supercontinent more than 45 million years ago. Similar now

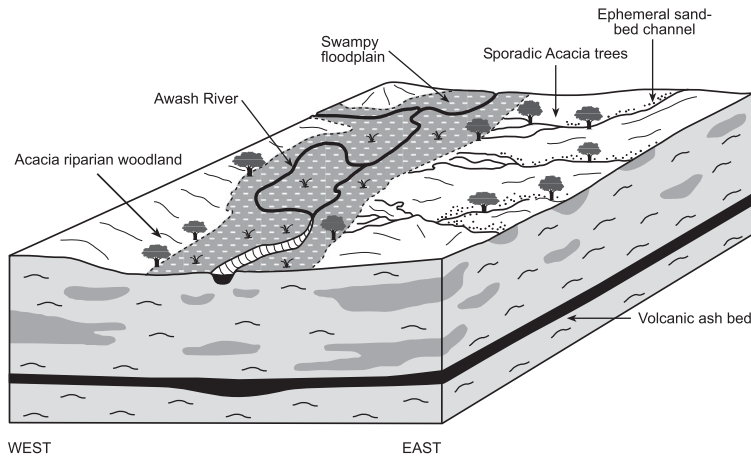


Figure 10.1. Pleistocene alluvium of the Awash River, southern Afar Rift, Ethiopia. (After Williams et al., 1986.)

defunct river valleys in the Namib and Mauretanian deserts are sometimes lined with beds of hard secondary calcium carbonate or calcrete (see [Chapter 15](#)), often rich in uranium (Netterberg, 1980; Dingle et al., 1983, p. 294). These calcretes represent former alluvial deposits that were subsequently cemented by groundwater rich in dissolved calcium carbonate – a legacy of past climatic changes.

The great sandstone plateaux of the eastern Sahara, such as the Gilf Kebir in hyper-arid south-eastern Libya and south-western Egypt, are deeply dissected by now inactive river valleys along their crenulated margins (Bagnold, 1933; Sandford, 1933; Peel, 1939; Peel, 1941), as are the more recently discovered sandstone plateaux in southern Libya to the north-east of Tibesti (Williams and Hall, 1965; Pesce, 1968; Griffin, 2006; Griffin, 2011). Peel (1966) and many later workers have emphasised the former efficacy of such fluvial erosion (e.g., Griffin, 2006; Griffin, 2011; Reid, 2009), which they consider strong evidence of formerly wetter or ‘pluvial’ climates (see [Chapter 12](#)). Carol Breed and her colleagues have made brilliant use of shuttle-imaging radar to map a whole series of late Cenozoic river channels in the eastern Sahara, some with associated prehistoric stone tool assemblages ranging typologically from Acheulian to Neolithic or from Early-Middle Pleistocene to Holocene in age (Breed et al., 1987; McHugh et al., 1988; McHugh et al., 1989).

Many of the great rivers that flow south from the Atlas Mountains, such as the Saoura, are flanked by alluvial terraces that contain pollen indicative of past changes in the vegetation cover in their headwaters (Beucher, 1971; Maley, 1980; Maley, 1981). The rivers that once flowed south from Tibesti into a much-expanded Lake Chad, or ‘Mega-Chad’, terminate in a delta dating from early to mid- Holocene (Schneider, 1967; Ergenzinger, 1968; Servant et al., 1969; Maley et al., 1970; Servant and Servant-Vildary, 1980; Drake and Bristow, 2006), when much of the Sahara was studded with small lakes (Faure, 1962; Hugot, 1962; Faure et al., 1963; Faure, 1966; Faure, 1969;



Figure 10.2. Alluvial terrace exposed on the flanks of a desert dune immediately east of the Aïr Mountains, south-central Sahara.

Hugot, 1977; Williams et al., 1987) joined by an integrated drainage network (Drake et al., 2011; Coulthard et al., 2013). The facies changes displayed by these river sediments can throw some light on past changes in flow regime. For example, the late Pleistocene alluvial terraces in narrow valleys draining Tibesti volcano (Hagedorn and Jäkel, 1969) and the eastern flanks of the northern Aïr Massif (Figure 10.2) (Williams, 1971; Williams, 1973b; Williams et al., 1987; Williams, 2008) consist of coarse sands and gravels at the base, fining upwards into finely laminated silts and clays, often containing unbroken freshwater gastropods (see Chapter 16). The gravel facies is indicative of episodic high-energy flow, while the fine-grained horizontally bedded silts and clays represent former flood-plain sediments laid down under a low-energy flow regime (Williams, 2008). However, we need to bear in mind that the rivers which flow into or, more rarely, across deserts are not always good indicators of local climatic conditions but may instead reflect environmental conditions in their headwaters.

Wasson (1996, p. 6) has pointed out that, although it is useful to compile global inventories of present-day river discharge and sediment load, the modern records are too short to provide robust insights into the sensitivity of river systems to climatic and land-use changes. For this we need the longer-term records provided by a more comprehensive analysis of river history (Macklin et al., 2012; Vita-Finzi, 2012a; Vita-Finzi, 2012b; Williams, 2012a). The aim of this chapter is therefore to consider the scope and limitations of using river sediments and fluvial landforms as indicators

Table 10.1. *Some attributes of fluvial systems in arid and in humid regions*

Humid Regions	Arid Regions
Abundant precipitation	Limited precipitation
Dense plant cover	Bare/sparse plant cover
Abundant shallow and deep groundwater	Limited shallow groundwater
Rapid, complete weathering	Slow, incomplete weathering
Deep cohesive slope mantles	Shallow, non-cohesive slope mantles
High infiltration, low run-off	Low infiltration, high run-off
High base flow	Low base flow
Regular perennial stream flow	Ephemeral/seasonal flow, flash floods
Low sediment yield (silt, clay)	High sediment yield (sand, gravel)
Suspension-load channel	Bed-load channel
Stable sinuous channel	Unstable shifting channel
Deep and narrow channel	Wide and shallow channel
Low stream gradient	Steep stream gradient
Low rate of flood-plain build-up	Rapid rates of local aggradation
External outlet to drainage basin	Often no outlet and internally drained

of past environmental change and to see whether unambiguous climatic signals can be discerned by studying the history of desert rivers. Before we do this, it is worth considering some of the more important differences between river systems in humid regions and those in deserts (Table 10.1).

10.2 Some attributes of desert river systems

De Martonne and Aufrère’s (1928) threefold classification of river systems as endoreic, exoreic and areic, mentioned in Chapter 4, remains a useful one, provided that we remain aware that river status may change over time as a result of tectonic, volcanic or climatic events (Frostick and Reid, 1987b; Vita-Finzi, 2012a; Vita-Finzi, 2012b). For example, an exoreic river system that once flowed to the sea may become blocked by tectonic uplift and diverted inland, thereby becoming endoreic. Climatic desiccation may then lead to disintegration of the drainage network, converting it from endoreic to areic.

An equally important distinction is that recognised recently between two quite different types of river system, one termed a *distributary fluvial system* (Weissmann et al., 2010) and the other the conventional axial drainage system with its main channel (Figures 10.3 and 10.4), levees, flood-plains, terraces (abandoned flood-plains: Figures 10.2 and 10.5) and back swamps. These latter systems have been investigated in detail by hydrologists and fluvial geomorphologists during the past fifty and more years (Baulig, 1950; Leopold et al., 1964; Gregory and Walling, 1973; Gregory, 1977; Schumm, 1977; Gregory et al., 1995; Anderson et al., 1996; Inam et al., 2007; Singh, 2007). The distinction between a distributive fluvial system and an



Figure 10.3. Ephemeral stream channel, Xinjiang Province, north-west China.



Figure 10.4. Ephemeral stream channel, Dire Dawa, southern Afar Desert.



Figure 10.5. Alluvial terrace, north of Aqaba, Jordan Desert.

axial fluvial system is important, because there is relatively little sediment deposition from axial river channels onto their flood-plains, in contrast to that associated with distributary fluvial systems (Weissmann et al., 2010). Leier et al. (2005) referred to what they termed *megafans*, which they considered to be diagnostic of river basins located in monsoonal or seasonally wet tropical regions with mountainous headwaters. In fact, a bigger sample of river basins (600) than that studied by Leier et al. (2005) indicates that megafans are included within the broader class of distributary fluvial systems, which span a wider range of climatic zones (Weissmann et al., 2010).

Another fluvial feature common in arid central Australia and elsewhere are ‘floodouts’, which may be defined as sites ‘where channelized flow ceases and floodwaters spill across adjacent alluvial surfaces’ (Tooth, 1999, p. 222). The ‘floodout zone’ is ‘that part of the [ephemeral stream] system where there is a marked reduction in channel capacity compared with reaches upstream and where overbank flows become increasingly important’ (op. cit., p. 222). Tooth identified four main agents responsible for the formation of floodouts, all involving barriers to flow (bedrock, eolian, hydrologic, alluvial). Of these, the most important is channel constriction upstream from Quaternary river terraces cemented by carbonate, iron or silica, and the burial of these terraces down valley by younger alluvium, with a decline in the size and capacity of the trunk stream channel, leading to overflow but without

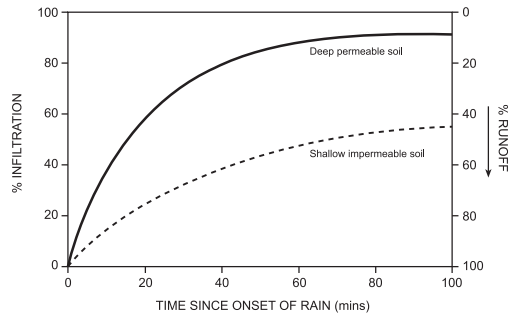


Figure 10.6. Run-off and infiltration associated with deep and shallow soils.

the development of distributary channels. He concluded that floodouts are neither alluvial fans nor terminal fans but a distinctive landform related genetically to floodplains.

Table 10.1 is a very simple and highly schematic summary of some of the more important differences between desert river systems and those located in well-vegetated humid regions, one of which (cited at the start of this chapter) was expressed succinctly by G.K. Gilbert (1877) more than a century ago. Where annual rainfall is sufficiently high, as in tropical and temperate humid regions, the soil surface along the valley sides has a dense plant cover, with deep, well-structured soils and abundant soil pores, so infiltration rates tend to be high and run-off tends to be relatively low (Figure 10.6). In contrast, under the sparse or absent plant cover of arid areas, the bare soil surfaces are prone to form impermeable surface crusts as a result of raindrop impact, so that infiltration rates into the often shallow soils are low and run-off rates are correspondingly high. As a result, desert streams are prone to flash floods and have typically erratic, ephemeral or highly seasonal flow regimes, in contrast to the regular perennial stream flow of humid regions, where the streams are maintained by high rates of base flow.

Because the valley sides in humid regions are protected from erosion by the dense plant cover, the amount of sediment supplied to the stream channels tends to be relatively low and often quite fine, again in contrast to desert rivers with their coarse sediment loads and high rates of sediment influx during the intense but infrequent storm events. In well-vegetated humid regions the streams tend to carry a sizeable suspension load of clay- and silt-sized particles and to be sinuous in plan view and often well-entrenched into their flood-plains. Desert streams, on the other hand, tend to be bed-load streams with steep, unstable, shifting channels. Obviously, there are exceptions to each of these generalisations. The highest rates of sediment yield in rivers come from both semi-arid and seasonally wet tropical regions (Douglas, 1967; Milliman, 1997). In effect, the seasonally wet tropics operate as semi-deserts during the dry season and as humid tropics during the wet season.

Table 10.2. *Erosion processes*

Agent	Erosion process	Sediment produced
Raindrop impact	Rainsplash	Colluvium
Overland flow	Slopewash	Colluvium
Sheet-floods	Slopewash	Colluvium
Slow mass movement	Soil and rock creep	Colluvium
Rapid mass movement	Landslides, mudflows	Colluvium
Rills and gullies	Channelled flow	Alluvium
Streams and rivers	Channelled flow	Alluvium
Frost and ice	Gelifraction, solifluction	Periglacial deposits
Glaciers, ice caps	Glacial erosion	Glacial deposits
Wind	Deflation, wind abrasion	Eolian sand, loess
Waves	Beach erosion, long-shore drift	Beach deposits
Groundwater	Leaching and solution	Precipitates

10.3 Sediment sources

In most desert rivers originating from upland regions, the bulk of the sediment comes from the mountainous headwaters, as for example in the Nile Basin, where the Ethiopian headwaters supply most of the load and much of the wet season discharge. The same situation holds true for large tropical rivers like the Amazon, in which roughly 90 per cent of the dissolved and suspended loads comes from 10 per cent of the basin area, namely the Andean headwaters (Gibbs, 1967; Meade, 2007). Given the importance of the headwaters as sources of sediment, we will begin with a brief survey of weathering and erosion processes in this critical sector. Even in deserts, erosion by running water is the most effective agent of geological erosion; wind erosion is limited to local undercutting of the softer rocks and sediments (Chapter 9, Figure 9.2). Desert dunes are themselves reworked sediments laid down initially by rivers or as beach deposits of former lakes.

Erosion is the detachment and transportation of earth materials. *Geological erosion* (or denudation) involves the wearing down and/or wearing back of upland areas to achieve eventually a surface of low relief in tectonically stable areas with a long history of weathering and erosion. Such erosion surfaces have been variously termed peneplains, peneplanes, pediplains, pediplanes or planation surfaces, all of which have in-built assumptions about how they might have formed (Davis, 1909; Davis, 1912; Penck, 1924; Penck, 1953). In order to avoid genetic connotations, it seems wiser simply to use the term *erosion surface*. *Accelerated erosion* is the destruction and removal of soil at a rate that is perceptibly faster than the geological rate characteristic of that region. For soil to form, there must be a rough balance between weathering and erosion. In seasonally wet tropical regions, soil formation may take thousands to tens of thousands of years (10^3 – 10^4 years), while soil loss may only take decades or centuries (10^1 – 10^2 years). Table 10.2 lists

some of the more common agents of erosion, which we now consider in more detail.

Raindrop impact is the initial agent of soil particle detachment. The kinetic energy (E_k) incorporated in falling rain is the product of total raindrop mass (m) and the square of the velocity (v) of the falling raindrops.

$$E_k = mv^2 \quad (10.1)$$

The momentum (M) of a falling drop at impact is the product of m and v .

$$M = mv \quad (10.2)$$

As might be expected, there is a linear relationship between the rainfall intensity of any given rainstorm and the total raindrop momentum embodied in that storm. The rate of detachment of soil particles is a function of total raindrop momentum (Williams, 1969a). At the moment of impact, some particles rebound upslope and some downslope. The steeper the slope, the higher the number of particles projected downslope. In any event, once run-off or overland flow occurs, detached soil particles will be washed downslope.

Desert storms are often very local and very intense. We might therefore expect high rates of soil particle detachment. This will only apply if the surface is not protected by a more or less continuous layer of stones or 'desert pavement' and, of course, if the surface is not bare rock. Surface rock creep is the slow movement downslope of surface rock fragments under the influence of gravity. The rate of movement is proportional to the sine of the angle of slope, regardless of whether the initial disturbance of the surface stones is triggered by frost or by raindrop impact.

A key factor limiting the impact of falling raindrops is the degree of plant cover or surface mulch, because these will absorb much or all of the momentum and kinetic energy of the falling drops. By way of example, on granite slopes of only 2 per cent gradient in the seasonally wet tropics of northern Australia, for the same unit momentum of falling rain, soil particle detachment was twenty to forty times more at the start of the wet season when the surface was bare than it was during the height of the rainy season when surface plant cover amounted to 30–40 per cent by area (Williams, 1969a).

One common effect of raindrop impact in arid areas is the creation of a thin surface crust formed of silt and clay particles. Such crusts are often only a few millimetres thick but can be many hundreds of times less permeable than the soil beneath them. The immediate consequence of such crusts is to reduce infiltration of water into the soil and produce a corresponding increase in surface run-off.

Where much of the surface consists of bare rock, run-off rates will be high. If, as is common in many semi-arid areas underlain by granite or similar rocks, the local relief consists of bare rocky hills, or *inselbergs*, rising above gently sloping plains, the foot-slopes around the *inselbergs* are often quite densely vegetated. This is

because run-off shed from the rocky hill slopes is absorbed by the aureole of sandy colluvial-alluvial sediment at the base of the granite hills, so that the soil moisture is enough to sustain the growth of perennial trees and shrubs. Another less obvious factor is subsurface lateral flow of water through the soil concentrated at the foot of the hill, leading to *eluviation*, or mechanical leaching, of the finer clay and silt particles from the weathered rock and soil and their deposition as *illuvial* clay particles several hundred metres further out in the plain at shallow depths. The higher clay content in these soils enhances their capacity to store water for plant growth. (The importance of *eluvial* and *illuvial* processes in soil formation is discussed in [Chapter 15](#).)

Run-off occurs during rain once the infiltration capacity of the soil is exceeded, so any excess rain is shed as surface run-off or overland flow. In the case of deep permeable soils, there may be little or no run-off unless rainfall intensities are very high. Run-off may at first be confined to shallow depressions, but if rainfall persists or is intense, the entire surface may become submerged beneath a sheet-flood. With the additional turbulence imparted by falling rain, run-off can be an efficient agent of surface erosion, the resulting slopewash deposits often showing weak, sub-horizontal planar bedding structures. The rate of erosion will vary with slope length, gradient and curvature, as well as with the nature of the surface (rocky, bare, vegetated) and type of soil. As a rule, sandy soils are more prone to erode under slopewash than are more cohesive clay soils.

In many instances, surface run-off is channelled into shallow linear depressions, or rills. Such rills have been termed first-order channels. Where two rills join, the channel below the confluence is considered a second-order channel and so on, until one is dealing with large drainage basins in which the axial channel may be a sixth-order channel or greater. In practice, it is a moot point as to what type of channel should be included, especially in the case of highly ephemeral or intermittent streams. Nevertheless, there are some well-established empirical relationships between stream order and other drainage basin attributes, such as sediment yield and discharge. Another measure is drainage density, or the basin area divided by the total stream length. High drainage densities are common in areas of severe gully or 'badland' erosion. Drainage density is high where soil permeability is low, as with many clay soils.

This description of raindrop impact, infiltration, run-off and rill erosion is highly generalised. The reality is more complex. Within desert landscapes, even subtle differences in microtopography and in plant type and distribution can exert a profound influence on local rates of infiltration ([Dunkerley, 2000](#)). Run-off itself is not a uniform process, because within a thin sheet of flowing water, there are concentrated threads of deeper and faster flow, which play an important role in sediment transport ([Dunkerley, 2004](#)). In addition, the definition of what a rainfall event is will vary with how the intervals between events are defined, so modelling the impact

of individual rainstorms can become problematic (Dunkerley, 2008a; Dunkerley, 2008b). However, this complexity is becoming better understood and appropriate techniques are being devised to monitor the nature and action of these processes (Dunkerley, 2010).

Two other processes of hill slope erosion strongly controlled by gravitational forces are slow mass movement (soil creep and surface rock creep) and rapid mass movement (landslides, mudflows and debris flows). There are few long-term studies of soil creep, but what data do exist suggest that volumetric rates of soil movement by soil creep may equal those from slopewash under the same climatic regime and on rocks of similar type, such as sandstone or granite (Williams, 1973a). Volumetric data on landslide activity in desert areas are sparse, but there are many accounts of mudflows and debris flows triggered by sudden downpours in arid and semi-arid areas. Since Archimedes' Principle pertains to all fluids, mudflows have a far greater capacity to transport large boulders than does an equivalent volume of less muddy water. Out on the plains surrounding Jebel Archenu ring-complex in the south-eastern Libyan Desert, there are large boulders up to 1 m in size that have quite probably been carried by former debris flows, the finer portions of which have long since been washed and blown away. A simple way to distinguish debris flow deposits from river alluvium is to examine whether or not the clasts or rock fragments within the deposits are in contact with each other (*clast-supported*) or are encased by finer sediment (*matrix-supported*). Alluvial deposits are generally clast-supported. However, many alluvial fan deposits (Figures 10.7 and 10.8) contain a mixture of both clast-supported and matrix-supported materials, indicating that they were formed by both debris flows and channel flows. Hill slopes are not the only suppliers of sediment to rivers. Another important source comes from bed and bank erosion in gullies, arroyos and ephemeral stream channels, to which we now turn.

10.4 Gullies, arroyos and ephemeral stream channels

In few parts of the desert world have ephemeral stream channels been investigated more thoroughly than they have in the American Southwest (Bryan, 1925a; Bryan, 1925b; Leopold and Miller, 1956; Schumm and Hadley, 1957; Bull, 1964a; Bull, 1964b; Lamarche, 1966; Leopold et al., 1966; Tuan, 1966; Haynes, 1968; Cooke and Reeves, 1976; Graf, 1979; Balling and Wells, 1980; Graf, 1982; Graf, 1983a; Graf, 1983b; Graf, 1987a; Bull, 1991; Schumm, 1991, pp. 108–119; Bull, 1997; Tucker et al., 2006). Three of the most comprehensive reviews of this topic are those by Cooke and Reeves (1976), Graf (1987a) and Bull (1997). The outstanding investigations in the American Southwest (summarised by Graf, 1987a) inspired a series of comparable studies in the Negev Desert of southern Israel (Schick, 1974; Schick et al., 1987; Laronne and Reid, 1993; Ya'ir and Lavee, 1976; Ya'ir and Lavee,



Figure 10.7. Alluvial fan, Negev Desert, Israel.



Figure 10.8. Exposed side view of alluvial fan, Negev Desert, Israel.

1982; Avni, 2005; Avni et al., 2006), as well as in semi-arid Australia (Wasson, 1976; Williams et al., 1991b; Dunkerley, 1992; Tooth, 1999; Tooth, 2000; Tooth and Nanson, 2000; Williams et al., 2001; Tooth and Nanson, 2004; Dunkerley, 2008c; Dunkerley, 2010; Glasby et al., 2010; Haberlah et al., 2010a; Haberlah et al., 2010b).

Bull (1997) drew a distinction between gullies and arroyos. He defined an arroyo as a ‘continuously entrenched stream channel in cohesive valley-floor alluvium’ (op. cit., p. 228). The length varied from 5 to 200 km, and the channel cross-section was typically a flat floor and near vertical banks. Unlike gullies, which are relatively small and exist for only a few years, arroyos may persist for more than a century. Bull also noted that ‘ephemeral streamflow is typical of many arroyos, but intermittent (occasional ground-water inflow) or perennial flow is common in the arroyos of large drainage basins’ (op. cit., p. 228).

One of the key questions relating to arroyos is why they sometimes deposit sediment along their beds and sometimes cut down. Many hypotheses have been advanced to account for arroyo incision, including climatic change and human activity. In fact, the reality is far more complex, as Cooke and Reeves (1976) have rigorously demonstrated. Their conceptual model of arroyo formation (op. cit., fig. 1.2) shows that more than thirty variables may be involved, so a search for single-cause explanations will generally prove fruitless.

Bull (1997) concurred with this conclusion and proposed a model based on changes in the balance between stream power and resistance to erosion. Bagnold (1966) defined stream power (w) as the rate of energy loss per unit length of stream, expressed as the product of tractive force (r) and velocity (v) per unit width of channel:

$$w = rv \quad (10.3)$$

Tractive force is the product of hydraulic radius (R) (i.e., the submerged channel cross section area divided by the wetted perimeter), slope (S) and the specific weight of the fluid (γ):

$$r = \gamma RS \quad (10.4)$$

Both stream power and sediment transport rate are proportional to stream velocity cubed (Schumm, 1977). Once stream power falls below a limiting threshold value, bank erosion and sediment transport will diminish, leading to local sedimentation (i.e., aggradation) within the ephemeral stream channel. Bull (1997) emphasised the importance of plant cover in minimising erosion and in promoting sedimentation within the ephemeral stream network, a conclusion confirmed by the work of Tucker et al. (2006) in the semi-arid rangelands of the Colorado High plains. Once incision occurred, the surface soil mantle would become drier, plants would die and run-off would be concentrated towards the arroyo headwalls, further enhancing channel entrenchment. Bull’s analysis of radiocarbon-dated Holocene arroyo bank

sections showing multiple episodes of degradation and aggradation suggested that arroyo incision could occur within less than 100 years, but more than 500 years might be needed for the complete aggradation of valley floors and incised stream channels within an ephemeral stream system. The lack of synchronism between episodes of incision in adjacent arroyo systems revealed that local hydrologic factors could outweigh the influence of regional drought and flood events, except in the case of extreme floods following sustained droughts in some of the larger ephemeral streams. An interesting by-product of the research into arroyos has been the realisation that well-meaning (and expensive) attempts at soil conservation were often either ineffective or unnecessary, depending on what stage of development the arroyo system had reached (Tuan, 1966; Bull, 1997).

The role of vegetation change on alluvial deposition in arid areas is in fact more complex than either Bull (1997) or Tucker et al. (2006) realised. Antinao and McDonald (2013) studied four localities at different elevations in the Mojave and northern Sonoran deserts. They found that the onset of deposition on alluvial fans began well before any changes in plant cover upstream, and could occur during several distinct combinations of vegetation change. They concluded that other factors probably controlled late Pleistocene and Holocene fan aggradation in this region, including local storm intensity and changes in the routes taken by water and sediment on hill slopes.

Vegetation also exerts a powerful control on channel patterns in ephemeral streams, most noticeably as one of the major causes of anabranching (Tooth and Nanson, 2000; Tooth and Nanson, 2004). In fact, over the past decade, many river channels that were once regarded as braided have increasingly been identified as anabranching channels, but the fundamental causes of anabranching have remained enigmatic. A detailed study of the Marshall River in arid central Australia has shed some useful light on this question (Tooth and Nanson, 2000). This work highlights the importance of one small local tree, the inland tea-tree (*Melaleuca glomerata*), in controlling the anabranching form of ephemeral stream channels in this arid region. The tea-trees first establish themselves sporadically within the sandy channels, forming obstacles below which sandy alluvium accumulates. Small ridges of sand develop within the channel bed, causing the diversion of flow around them. As the ridges become longer and join with other mid-channel ridges, a long vegetated ridge develops and divides the channel into anabranches. This has the effect of decreasing resistance to flow and allows efficient movement of water and sediment within the anabranch channels. Anabranching can thus be seen as a means by which ephemeral stream channels can maintain the flow of water and sediment in situations where the river cannot increase its gradient. Unlike braided channels, where the mid-channel bars are unstable and ephemeral features, the vegetated ridges between anabranch channels are stable and relatively permanent landforms. This study very clearly demonstrates the important effects of vegetation on fluvial processes in arid areas.

10.5 Suspension load, bed load and river metamorphosis

River channel patterns have been the focus of considerable study (Fabre, 1797; Leopold and Wolman, 1957; Leopold et al., 1964; Gregory and Walling, 1973; Gregory, 1977; Schumm, 1977; Anderson et al., 1996). Three main patterns were identified early on – meandering, braided and straight. To these we may add *anabranching*, *anastomosing* and *tributary* channels. Meandering channels are sinuous in plan form, relatively deep and narrow, with cohesive beds and banks and gentle gradients. They carry most of their load in suspension. Braided channels, on the other hand, tend to be wide and shallow, with non-cohesive beds and banks, frequent mid-channel bars, and are relatively steep. In contrast to the meandering suspension-load channels, braided channels typically transport most of their sediment load along the channel bed and are therefore termed bed-load, or traction-load, channels. Mixed-load channels, as the name implies, transport roughly equal proportions of bed-load and suspended load. An anabranching channel is one that leaves the parent channel at some point along its lower course to rejoin it tens of kilometres further downstream (Nanson and Knighton, 1996). Such channels are common in the lower White Nile Valley of central Sudan and in the lower Darling River in semi-arid New South Wales, Australia. Anastomosis is an extreme form of anabranching, involving multiple channels. The Channel Country of south-west Queensland is a classic example of anastomosing channels during times of extreme flooding; so too are the Sudd Swamps of South Sudan and the Okavango ‘inland delta’ during flood. The ‘inland delta’ of the Niger is another example, complicated further by partially submerged dunes. Tributary channel patterns are characteristic of both deltas and alluvial fans and are diagnostic of all *tributary fluvial systems* (Weissmann, 2010), regardless of regional climate.

The early empirical studies of small American rivers (Leopold and Wolman, 1957; Leopold et al., 1964; Schumm, 1977) demonstrated that river channel pattern and form were closely related to the type of sediment transported, and changed in response to changes in sediment supply, sediment calibre and rainfall regime. Changes in the amount and particle size of sediment transported are controlled by events upstream, especially changes in vegetation type and cover, as noted by Bull (1997). Such changes in plant cover and precipitation will in turn control the ratio of load to discharge within the stream channel, as well as the calibre of material ferried from hill slope to river channel. Where the vegetation cover in the upper reaches of desert rivers is relatively sparse and the rainfall regime is prone to sporadic, highly intense and often very local downpours, the rivers will most commonly display a braided channel pattern and carry a sizeable traction load of coarse and non-cohesive cobbles and even quite large boulders. Because stream power is roughly proportional to velocity cubed and particle size carried in traction is proportional to about the sixth power of stream velocity, desert flash floods are highly efficient agents of erosion. The ephemeral stream channels change course frequently once they have left the confines of narrow

mountain valleys and spread out across the piedmont or foot-slopes, rapidly losing power and depositing poorly sorted coarse debris onto a growing alluvial fan.

Where the plant cover in the headwaters is dense and the rainfall regime is less erratic, as in the seasonally wet tropics and many semi-arid areas, rock weathering will be enhanced and deep soil mantles will cover many of the hill slopes and valley sides, contributing fine sediments to the river channels from slopewash and soil creep. The recipient stream channels will be suspension-load channels with deep and narrow channel cross sections, cohesive beds and banks, and a meandering stream pattern. Small ephemeral stream channels can change from braided to meandering within a few years or decades in response to local hydrologic changes triggered by changes in plant cover. Larger river channels can change from one type of channel to another during longer intervals of time, depending on changes in stream discharge and sediment load, a process described by Schumm (1969) as 'river metamorphosis'. We saw in [Section 10.4](#) that both stream power and sediment transport rate are proportional to stream velocity cubed. Once stream power falls below a limiting threshold value, bank erosion and sediment transport will diminish, leading to a change in channel pattern from braided to meandering.

10.6 Quaternary paleochannels in semi-arid south-east Australia

The largest rivers in Australia today rise in the Eastern Highlands and flow west or south-west across semi-arid alluvial plains. These rivers depend primarily on summer rainfall in the north and spring snow-melt in the south. A series of former river channels are clearly evident across the semi-arid Riverine Plain in south-east Australia ([Figure 10.9](#)). In a classic study of the Murrumbidgee River and its associated former channels, Schumm (1968) examined borehole data and channel dimensions and sought to explain how the changes in stream pattern and channel size reflected what he termed 'river adjustment to altered hydrologic regimen'.

Schumm observed that there were two distinct types of channel with quite different types of sediments within them. He adopted the names *ancestral stream* and *prior stream* that had been used by earlier workers for these two channel types. The ancestral stream channels were sinuous with meander wavelengths several times those of the present meandering channel, and were filled with mainly fine sediment, consistent with their sinuous channel pattern. In contrast, the prior stream channels contained a coarser channel fill and were linear in plan, with wide, relatively straight channels. Schumm concluded that the sinuous ancestral channels were suspension-load channels formed when the climate was wetter and bankfull discharge was several times greater than they are today. He proposed that the prior stream channels developed under a more seasonal flow regime, with sporadic episodes of very high discharge coming from more sparsely vegetated headwaters.

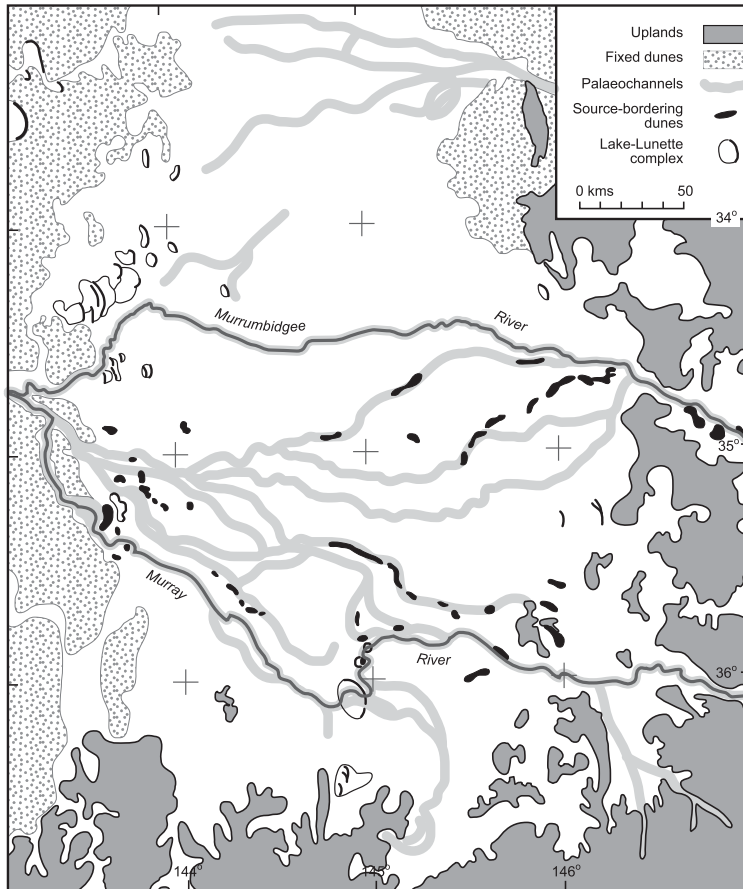


Figure 10.9. Murrumbidgee paleochannels, Australia. (Compiled for the author by the late Jim Baxter, 1986.)

Later workers built on this pioneering research and, in particular, were finally able to obtain radiocarbon ages for charcoal fragments and logs within the channel fill sediments and so provide a coherent chronology for the late Quaternary paleochannels of the Riverine Plain, in the process also dating episodes of neotectonic activity within the catchments (Bowler, 1978a). Before and during the Last Glacial Maximum (21 ± 2 ka), the rivers flowing west from the Eastern Highlands were ferrying a coarse load of sand and gravel across the alluvial plains, with widespread alluviation until about 15 ka. Aggradation was concentrated around 30–25, 20–18 and 18–14 ka (TL, OSL and ^{14}C ages) (Page et al., 1991; Page and Nanson, 1996; Page et al., 2001). As noted earlier by Schumm (1968), the meander wavelengths and channel widths indicated greater fluvial discharge, at least seasonally, before and during the LGM (Bowler, 1978b; Page et al., 1991; Page et al., 1996). The rivers in northern Victoria

and southern New South Wales had much higher late Pleistocene flood discharges and bed loads than they do today.

There is some evidence of a link between variations in river discharge and sediment load and short-lived phases of glacial and periglacial activity in their headwaters in the Eastern Highlands of south-east Australia (Williams et al., 2009b). The three youngest glacial advances in the semi-arid Snowy Mountains of south-east Australia have yielded ^{10}Be cosmogenic nuclide ages of 32 ± 2.5 , 19.1 ± 1.6 and 16.8 ± 1.4 ka (Barrows et al., 2001; Barrows et al., 2002). Periglacial deposits range in age between 23 and 16 ka, with a weighted average age of 21.9 ± 0.5 ka (Barrows et al., 2004). Episodes of at least seasonally very high river discharge in the Murray, Murrumbidgee, Lachlan and other major rivers draining the south-east uplands have been dated by ^{14}C , TL and OSL to approximately 35–25 and 20–14 ka (Bowler, 1978a; Page et al., 1991; Page et al., 1994; Page and Nanson, 1996; Page et al., 1996; Ogden et al., 2001; Page et al., 2001; Bowler et al., 2006). Reinfelds et al. (2014) have shown that in the Snowy Mountains today, run-off decreases by 17 per cent for every 1°C increase in annual temperature and proposed that lower LGM temperatures could have more than doubled run-off rates in that region.

The onset of the modern flow regime in the Riverine Plain of south-east Australia appears to be somewhat younger than the 16 ka start of deglaciation in the Snowy Mountains (Ogden et al., 2001; Barrows et al., 2001). No doubt an extensive winter snow cover persisted in the headwaters well after the valley glaciers had melted, contributing to high rates of seasonal river discharge. Late Pleistocene periglacial solifluction deposits are widespread in the uplands of south-east Australia (Bowler et al., 1976), and these would have provided an ample supply of coarse debris to streams during the spring snow-melt season.

Further north, beyond the limits of glacial and periglacial action, the sparse vegetation cover characteristic of the cold, dry late Pleistocene climate would also have been conducive to initially high rates of run-off and a relatively high load of coarse sediment, reflected in large stream channel dimensions out on the alluvial plains west of the uplands (Williams, 1984e; Williams, 2000b; Williams, 2001a; Williams, 2001b). As the climate became warmer, the plant cover became denser and soils began to develop, leading to a change from traction-load to mixed load to suspension-load channels. With the decline in rainfall during the past 5,000 years, the previously wide meandering channels became progressively smaller, with shorter meander wavelengths, culminating in the modern ‘underfit’ rivers, dwarfed by their late Pleistocene and early Holocene ancestors. Much of the evidence for late Quaternary climate change in south-east Australia comes not so much from the rivers themselves as from pollen analysis and studies of the lakes in this region (Williams et al., 2009b).

Fried (1993) suggested that the suspension load in the late Quaternary Riverine channels could have come from wind-blown dust, so that the large meanders could reflect the dust input rather than hydrologic changes in the headwaters. Later workers

have not pursued this idea, but it is not without merit (see Wasson, 1982). In the semi-arid Flinders Ranges of South Australia, the fine-grained late Pleistocene valley fills represent reworked eolian dust or loess mantles (Williams et al., 2001; Chor et al., 2003; Williams and Nitschke, 2005; Williams et al., 2006a; Williams and Adamson, 2008; Haberlah et al., 2010a; Haberlah et al., 2010b). Similar valley fills derived from reworked loess are a feature of the Matmata limestone uplands in Tunisia (Coudé-Gaussen et al., 1987), the Namib piedmont valleys (Eitel et al., 2001; Heine and Heine, 2002; Eitel et al., 2005) and the wadis within the Sinai Desert (Rögner et al., 2004).

A word of caution is needed here in relation to changes from braided to meandering stream patterns. It is widely assumed in much of the geomorphic literature relating to meandering and braided river channels that there are distinct hydrologic thresholds leading from one state to another. This is a view challenged by Tooth and Nanson (2004) as a result of their work on the Plenty and Marshall rivers in arid central Australia. Both rivers flow close to one another on roughly parallel courses in one sector of 70 km and have similar gradients and local climates. Despite this, and contrary to accepted geomorphic theory, they display very different channel patterns and cross-sectional forms. The Plenty River in the reach studied is a single-thread, low-sinuosity channel up to 2 m deep and 1,200 m wide. In strong contrast, the Marshall River has numerous narrow anabranches, usually less than 60 m wide, separated by vegetated ridges and broader islands. Arising from their field investigations, Tooth and Nanson (2004) concluded that there were two main reasons for this difference, both of which are quite subtle. The Marshall River carries a slightly coarser bed load and receives occasional tributaries, enabling tree growth and accumulation of sediment immediately upstream of these trees, leading to anabranch formation on either side of the obstructions. Hence, minor differences in sediment and water inflow to the trunk stream channel can lead to major changes in stream channel morphology.

10.7 Quaternary Blue Nile paleochannels on the Gezira plain, semi-arid Sudan

In this section, we focus in some detail on the Nile, not only because it has been intensively studied over many decades (Lombardini, 1865a; Lombardini, 1865b; Willcocks, 1904; Lyons, 1906; Lawson, 1927; Hurst and Philips, 1931; Hurst and Phillips, 1938; Hurst, 1952; Williams and Faure, 1980; Williams and Adamson, 1982; Hassan, 1981; Said, 1993; Said, 1997; Williams et al., 2000; Woodward et al., 2001; Woodward et al., 2007; Williams, 2009b; Williams et al., 2010b; Williams, 2012a), but especially because it illustrates very nicely how different lines of evidence can be used to construct a coherent history of river response to environmental change in arid areas.

The Nile is the longest river in the world and carries a sediment load of about 100 million tonnes/year, most of which comes from its Ethiopian headwaters, as Herodotus correctly surmised some 2,500 years ago. The two major Ethiopian

tributaries are the Blue Nile and the Atbara. The Blue Nile joins the White Nile at Khartoum to form the Main Nile, or Saharan Nile, which then flows north into the eastern Sahara Desert to be joined by the Atbara 320 km downstream, after which the Nile has a waterless journey of 2,689 km until it debouches into the Mediterranean some 256 km north of Cairo. The Atbara also rises in the highlands of Ethiopia not far from the source of the Blue Nile but pursues a more northerly course through more arid terrain, losing its water in the desert during the dry winter months, it failed to reach the main Nile until construction of the dam at Khashm el Girba.

The Gezira is a clay-mantled, low-angle alluvial fan roughly 230 km in radius and 40,000 km² in area. It is bounded to the east by the present Blue Nile, to the west by the White Nile and to the south by the Managil Ridge, and is traversed by a series of sandy paleochannels that originate between the towns of Sennar and Wad Medani on the present-day Blue Nile (Figure 10.10). The Blue Nile fan consists of a veneer of dark cracking clay one to five metres in thickness which mantles alluvial sands and gravels with very large cross-beds indicative of very high-energy flow (Williams, 2012a). Tothill (1946; 1948) was the first to show that the Gezira clays were alluvial and not formed by wind-blown dust, and considered that they had been laid down in early Holocene times as a result of seasonal flooding from the Blue Nile. He based his conclusions on the presence of aquatic gastropods (see Chapter 16) in the upper two metres of Gezira clay and on the presence of pyroxene and other heavy minerals indicative of a volcanic (i.e., Ethiopian) provenance for the alluvial clays. Williams (1966) proposed that the Gezira clays had been deposited by seasonal floods from distributary channels of the Blue Nile that radiated across the Gezira alluvial fan. He also provided the first radiocarbon ages for two sites east of the White Nile, showing that these White Nile alluvial clays had been deposited during the terminal Pleistocene and early Holocene.

Williams and Adamson (1974; 1980; 1982), Adamson et al. (1980; 1982) and Williams et al. (1982; 2000) carried out a program of comprehensive radiocarbon dating of gastropod and other shells from Blue and White Nile alluvium. These authors were able to show that there had been a series of Holocene phases of diminishing high flood levels in both rivers, with the last moist phase at around 2 ka associated with a unique form of pottery manufacture using the swamp-dwelling sponge *Eunapius nitens* as temper (Adamson et al., 1987a). Later work by Talbot et al. (2000) and by Williams et al. (2006c), based in part on the use of strontium analyses (see Chapter 7, Figure 7.1), showed that the abrupt return of the summer monsoon in the Ugandan headwaters of the White Nile around 14.5 ka was followed by widespread flooding and clay deposition in the lower White Nile Valley. This major flood event caused erosion of many of the sand dunes in the lower White Nile Valley (Williams, 2009b) and was followed by Blue and White Nile incision and progressively reduced flooding until aridity set in around 4,500 years ago (Williams et al., 2010b).

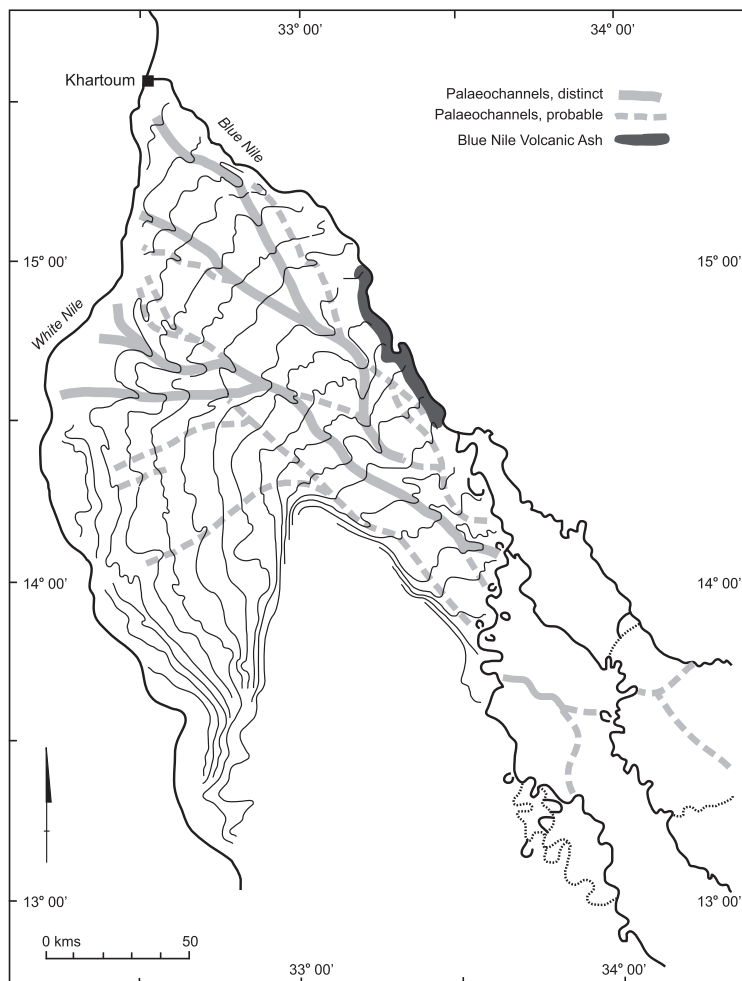


Figure 10.10. Blue Nile paleochannels, central Sudan. (After Williams, 2009b.)

The first attempt at luminescence dating of White Nile alluvium was on sediment cores collected from a 6 m deep trench dug near the village of Esh Shawal, 300 km upstream from the Blue and White Nile confluence at Khartoum (Figure 10.5). The luminescence ages revealed that White Nile alluvium dated back to at least 240 ka (Williams et al., 2003), consistent with independent but similar estimates for the inception of Lake Victoria in the Ugandan headwaters (Talbot and Williams, 2009; Williams and Talbot, 2009).

A series of sinuous Blue Nile paleochannels are very clearly visible on air photographs of the north-west Gezira (Figure 10.11). They continue towards the White Nile and disappear beneath the dunes located between Jebel Aulia in the north and Naima

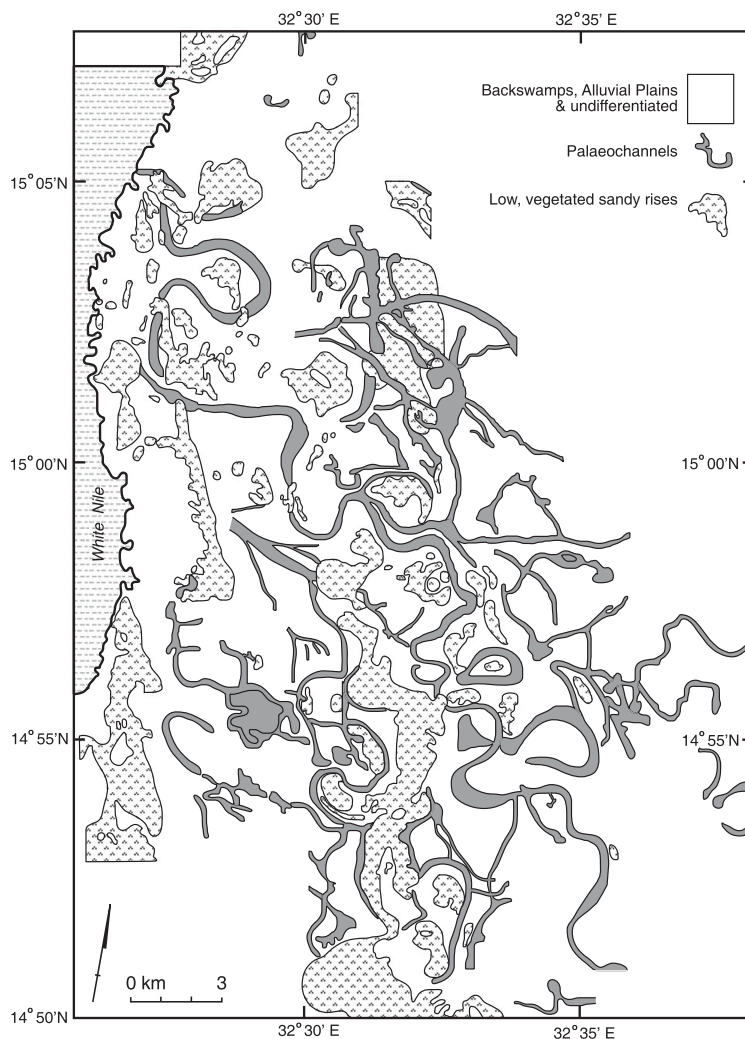


Figure 10.11. Late Pleistocene meanders east of the White Nile, central Sudan. (After Williams, 2009b.)

in the south. The channel fills consist of a clay layer over fine to medium sands. The sands have OSL ages indicating a prolonged phase of fluvial sand entrainment and deposition in this area between 100 and 70 ka, and the clays have OSL ages between 76 and 50 ka, suggesting that these channels remained active for some 50,000 years in this locality.

The late Quaternary palaeochannels that radiate across the Gezira carried a bed load of sand and fine gravel from the volcanic uplands of Ethiopia. The heavy mineral suite of the source-bordering dunes and sandy point-bars associated with these channels is virtually identical to the heavy mineral assemblage from channel sands collected by the

author from the bottom of the Blue Nile gorge near Debre Markos in the highlands of Ethiopia (Williams and Adamson, 1973). ^{14}C ages obtained previously on oyster and gastropod shells, charcoal and carbonate samples within the paleochannel sediments indicated that the Blue Nile paleochannels were at least seasonally active between >40 ka and 8 ka to 5 ka (Adamson et al., 1982). Current work has extended the age of these channels back to at least 100 ka. The abandonment of the channels was a direct result of incision by the main Blue Nile channel, which began at least 8,000 years ago (Arkell, 1949; Arkell, 1953; Adamson et al., 1982; Williams et al., 1982). This incision effectively beheaded the distributary channels and deprived them progressively of their flood discharge. As the Nile cut down into its former floodplain, a series of shallow drainage channels remained seasonally active and finally dried out.

The final stages of infilling of these paleochannels and adjacent floodplains involved deposition of a thin layer of dark grey-brown clay. The fining-upwards sequence reflects a change from the transport of pale yellow medium and coarse quartz sands and fine quartz and carbonate gravels to grey-brown silty clays, sandy clays and clays in the upper 50–150 cm. Clay deposition in the back-swamps and flood-plains of these channels dwindled and finally ceased about 5,000 years ago, when the seasonally flooded swampy plains gave way first to acacia-tall grass savanna and finally to semi-desert steppe. The fossil snail fauna within the upper two metres of Holocene Gezira clay shows a progressive change from permanent water species to semi-aquatic species with lungs and gills to land snails (Tothill, 1946; Tothill, 1948; Williams et al., 1982). One of these, the large land snail *Limicolaria flammata*, today inhabits the acacia-tall grass savanna region to the south of Sennar, where the annual rainfall is at least 450–500 mm, in contrast to the 175 mm that now falls at Khartoum. *Limicolaria* was at its most widespread in this region around 5.2 ka.

10.8 A late Pleistocene and early Holocene depositional model for the Blue Nile

Thirty years earlier, when far fewer radiometric ages were available for this region, Adamson et al. (1980) and Williams and Adamson (1980) proposed a simple depositional model linked to climate and plant cover to account for these changes. During cold, dry glacial intervals, the headwaters of major Ethiopian rivers would be sparsely vegetated, hill slope erosion would be accelerated and rivers would become highly seasonal, low-sinuosity, bed-load streams which carried and deposited large volumes of poorly sorted gravels and sands (Figure 10.12). Conversely, with a return to warm, wet conditions and re-establishment of a dense plant cover in the headwaters, we should see a change to high-sinuosity, suspended-load streams that carried and deposited silts and clays (Figure 10.13). A fining-upwards alluvial sequence from coarse basal gravels through sands to horizontally bedded silts and clays is thus a predictable outcome of a change from a bed-load to suspended-load regime, related to a change

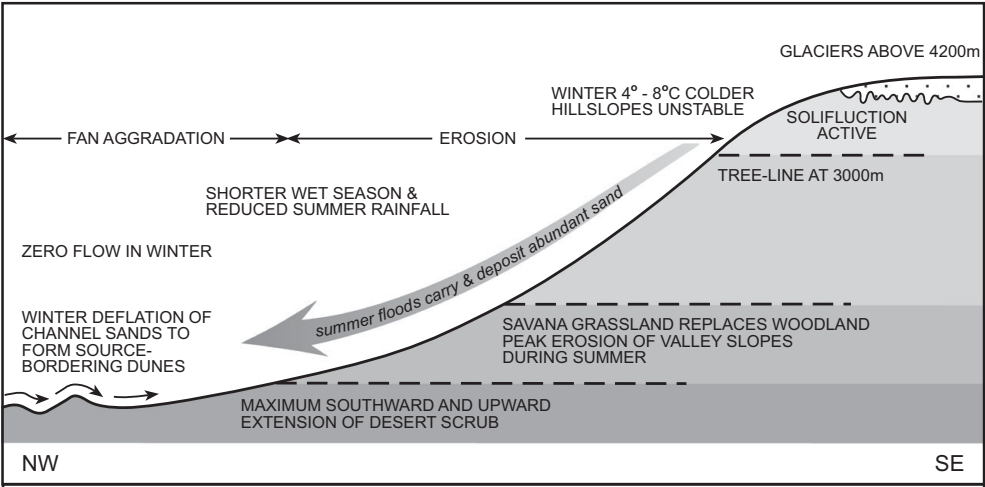


Figure 10.12. The late Pleistocene Blue Nile. (After Williams, 2012a.)

from glacial aridity (Table 10.3a) to interglacial and postglacial climatic amelioration (Table 10.3b). The precise timing of the last glaciation in the Ethiopian Highlands is still being investigated. Osmaston et al. (2005) considered that up to 180 km² of the Bale Mountains of Ethiopia could have been glaciated at this time, with a central ice cap of at least 30 km². Glacial moraines and periglacial deposits in the Semien Mountains near the sources of the Tekezze and Blue Nile/Abbai rivers are presently being dated using cosmogenic nuclides. The few available radiocarbon ages point to colder LGM conditions (4–8°C cooler), with a lowering of the upper tree-line by about 1,000 m during the LGM (Williams et al., 1978; Hurni, 1982).

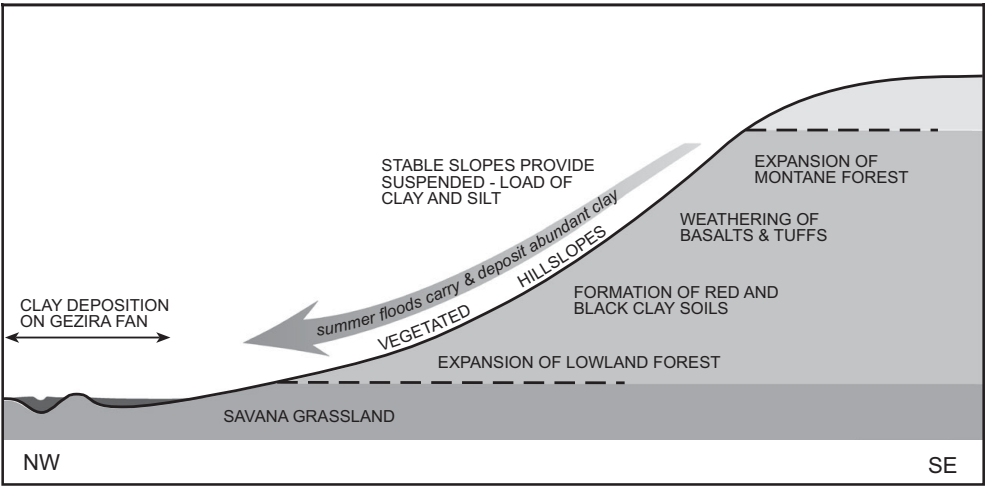


Figure 10.13. The early Holocene Blue Nile. (After Williams, 2012a.)

Table 10.3a. *The Blue Nile at 21–18 ka*

-
-
- Less summer rain
 - Winters 4–8°C cooler
 - Glaciers above 4,200 m
 - Slopes unstable down to 3,000 m, resulting in coarse debris
 - Tree-line 1,000 m lower in headwaters
 - Savanna/desert ecotone higher
 - High seasonal run-off, resulting in high peak flows but reduced annual discharge
 - Stream channel braided with sand and gravel point-bars
 - Winter deflation of channel sands, resulting in source-bordering dunes
-
-

The Blue Nile has cut down at least 10 m since 15 ka and at least 4 m since 9 ka, with concomitant incision by the White Nile amounting to 4 m since 15 ka and at least 2 m since 9 ka (Arkell, 1949; Arkell, 1953; Williams and Adamson, 1980; Williams et al., 2000). Such incision would help drain previously swampy flood-plains, freeing them for cultivation by Neolithic farmers. A number of episodes of high Nile floods occurred during this time and are reflected in the presence of cross-bedded sands and shell-bearing clays above the maximum unregulated flood levels. Thanks to a very gentle flood gradient (1:100 000), the post-LGM flood deposits in the lower White Nile Valley are well-preserved. Calibrated ^{14}C ages obtained on freshwater gastropod and amphibious *Pila* shells and fish bones show high White Nile flood levels around 14.7–13.1, 9.7–9.0, 7.9–7.6, 6.3 and 3.2–2.8 ka. The less complete Blue Nile record shows very high flood levels around 13.9–13.2, 8.6, 7.7 and 6.3 ka (Williams, 2009b).

Mayewski et al. (2004) synthesised the results from some fifty globally distributed paleoclimate records spanning the time interval from 11.5 ka to the present. They identified six significant periods of rapid climate change at 9–8, 6–5, 4.2–3.8, 1.2–1.0 and 0.6–0.15 ka, the first five of which coincided with polar cooling and tropical aridity. The intervals in between were wetter in the tropics and, allowing for dating

Table 10.3b. *The Blue Nile at 15–14 ka*

-
-
- High summer rainfall, resulting in a longer wet season
 - Winters 4–8°C warmer
 - Slopes vegetated and stable above and below 3,000 m
 - Tree-line 1,000 m higher in the headwaters
 - Expansion of lowland savanna into former semi-desert
 - Weathering of tuffs and basalts and formation of clay soils in uplands, resulting in a suspension load of silt and clay
 - Higher annual discharge and greater base-flow, resulting in attenuated peak floods
 - Perennial flow and prolonged, widespread flooding by sinuous (and straight) distributary channels
-
-

errors, tally reasonably well with the intervals of high Blue and White Nile floods identified here. At the site of Erkowit in the Red Sea Hills (Mawson and Williams, 1984), there is evidence of permanent stream flow around 1.8–1.6 ka, coinciding with high White Nile flows but not as yet evident in the much more incomplete Blue Nile sedimentary record.

10.9 East Mediterranean marine sediment core records of Quaternary Nile flow

During phases of very high Nile flow, clastic muds rich in continental organic matter and highly organic sapropels accumulated on the floor of the eastern Mediterranean (Rossignol-Strick et al., 1982; Rossignol-Strick, 1985; Lourens et al., 1996; Rossignol-Strick, 1999; Mercone et al., 2001; Krom et al., 2002; Larrasoana et al., 2003; Ducassou et al., 2008; Ducassou et al., 2009; Revel, 2010). Flood deposits exposed in trenches dug east of the present White Nile near Esh Shawal village 300 km south of Khartoum (Williams et al., 2003) show episodes of middle to late Pleistocene high flow which, within the limits of the dating errors, coincide with sapropel units S8 (217 ka), S7 (195 ka) and S6 (172 ka) (Lourens et al., 1996). Sapropel 5 (124 ka) was synchronous with major flooding in the White Nile Valley and with a prolonged wet phase at around 125 ka at Kharga Oasis in the Western Desert of Egypt (Kieniewicz and Smith, 2007). Recently dated high flood deposits on the main Nile are roughly coeval with sapropel units S6 (172 ka) and S3 (81 ka) (Williams et al., 2010b).

The most recent sapropel S1 in the eastern Mediterranean is a composite unit, with ages of 13.7–12.4 ka near the base and 9.9–8.9 ka near the top (Williams et al., 2010b). The gap in the S1 record may coincide with the arid phase seen in other parts of Africa coinciding with the Younger Dryas (around 12.5–11.5 ka). Higgs et al. (1994) considered that formation of sapropel S1 may have ended as recently as 5 ka, which is also when the Nile deep-sea turbidite system became inactive as a result of reduced sediment discharge from that river (Ducassou et al., 2009). The interval from around 13.7 to 8.9 ka and locally up to 5 ka also coincides with a time when freshwater lakes were widespread in hollows between the White Nile dunes (Williams and Nottage, 2006), as well as west of the Nile and in the eastern Sahara (Williams et al., 1974; Williams and Faure, 1980; Pachur et al., 1990; Pachur and Hoelzmann, 1991; Ayliffe et al., 1996; Pachur and Wünnemann, 1996; Hassan, 1997; Pachur and Altmann, 1997; Pachur and Hoelzmann, 2000; Hassan et al., 2001; Hoelzmann et al., 2004; Williams et al., 2010a; Williams and Jacobsen, 2011) and when the White Nile attained flood levels up to 3 m above its modern unregulated flood level.

Where independently dated comparisons exist between sapropel formation and Nile floods, they point to synchronism between sapropel accumulation and times of higher Nile flow, indicative of a stronger summer monsoon at these times. Although

the sapropel record in the eastern Mediterranean is incomplete, with some evidence of complete removal of sapropels by post-depositional oxidation (Higgs et al., 1994), it is a longer and more complete record than that presently available on land, and so can serve as a useful surrogate record for Nile floods and phases of enhanced summer monsoon precipitation.

10.10 Nile Delta records of Holocene fluctuations in Nile flow

Marriner et al. (2012) carried out a rigorous statistical comparison between mean rates of Nile deltaic sedimentation during the past 8,000 years (8 ka) and well-dated proxy records of climatic change in sites as far away as the Cariaco Basin off Venezuela, speleothems from Oman and China, a detailed lake diatom record from Ethiopia and Neolithic pastoral sites from the eastern Sahara. Two main conclusions emerged from their analysis. One, evident in high rates of deltaic sedimentation, confirmed that the early to mid-Holocene moister interval was associated with a northward displacement of the Intertropical Convergence Zone (ITCZ). Desiccation after 5.5 ka reflected a southward displacement of the ITCZ and a progressive decline in deltaic deposition. Nile Delta accretion during the Holocene depended on suspended sediment inputs from upstream, and these in turn were related to changes in monsoon strength controlled ultimately by the earth's orbital geometry.

The second significant conclusion concerns the link between El Niño-Southern Oscillation (ENSO) events, Nile flow, Nile sediment discharge and regional hydro-climatic changes at submillennial time scales. This link has long been very well-established for the time of instrumental records (see [Chapter 23](#)) but had not been shown explicitly for the Holocene. One tantalizing suggestion is that a weak statistical correlation between Nile deltaic sedimentation and proxy climate records during the interval 4.6–2.5 ka may reflect increasing human impacts in the Nile Valley, triggering changes in deltaic deposition.

10.11 River channel incision and deposition

It can be argued that the Blue Nile depositional model illustrated in [Figures 10.12 and 10.13](#) is based on an unproven assumption, namely, glacial aridity. After all, other workers had used the evidence afforded by late Pleistocene Nile sands and gravels flanking the Nile in northern Sudan and southern Egypt to argue for greater fluvial competence and consequently higher discharge and more pluvial glacial conditions (Butzer and Hansen, 1968). The contentious issue of glacial aridity versus glacial pluvial is discussed in [Chapter 12](#). Suffice to say here that the inference by Adamson et al. (1980) that the late Pleistocene was a time of greater aridity in the Nile headwaters was based on the fact that during the Last Glacial Maximum, lake levels in Ethiopia were low (Gasse, 1975; Gasse, 2000a; Gasse, 2000b), as they were in Kenya

(Butzer et al., 1972) and Uganda (Livingstone, 1980). In addition, as we saw in Chapter 8, Saharan desert dunes were active up to 800 km south of their present limits (Mainguet and Canon, 1976; Mainguet et al., 1980; Talbot, 1980; Swezey, 2001). The White Nile, deprived of the run-off from its headwaters by the closure of the Ugandan lakes, dried out in the winter months, during which sand dunes migrated across its former bed (Williams, 2009b).

To sum up, the late Pleistocene Blue Nile and Atbara rivers were highly seasonal, bed-load streams that, together with their tributaries, ferried and deposited vast quantities of poorly sorted sands and gravels in central Sudan and southern Egypt (Williams et al., 2010a). With the return of the summer monsoon around 17 ka, strengthening at 15 ka (Williams et al., 2006c), run-off increased in the Ethiopian headwaters and Lake Tana overflowed once more (Lamb et al., 2007). From around 15 ka until around 7.5 ka and perhaps slightly later (Williams, 2009b), the Holocene Blue Nile was depositing clays across the low-angle Gezira alluvial fan in the central Sudan. Thereafter, it began to incise, terminating its upward-fining depositional cycle.

A similar pattern of widespread deposition of late Pleistocene sand and gravel, followed by terminal Pleistocene to early Holocene fine-grained alluviation culminating in vertical river entrenchment has been documented for the Son and Belan rivers in semi-arid north-central India (Williams and Clarke, 1984; Williams et al., 2006b; Gibling et al., 2008), as well as in the subhumid to semi-arid Murray and Murrumbidgee river basin in south-eastern Australia (Bowler, 1978a; Bowler, 1978b; Page et al., 1991; Page et al., 1996). It thus appears that rivers in semi-arid catchments are sensitive to changes in plant cover, whether they were once glaciated or not. A substantial reduction in vegetation cover in their headwaters is conducive to a bed-load regime, reverting to a suspension-load regime once the plant cover has been restored and a soil cover has been widely established in the headwaters. In essence, in the absence of any eustatic, isostatic or other tectonic causes of changes in base level, a river will tend to aggrade its valley when the ratio of load to discharge is high and to degrade its valley when the ratio of load to discharge is low. However, care is needed to avoid falling into the trap of circular argument in which a given type of climate (wetter, drier, transitional from wet to dry or dry to wet) is inferred from the presence of a river deposit and the inferred climate is then used to account for the existence of the same deposit. Some independent check on the purely fluvial evidence is therefore necessary when seeking to reconstruct climatic changes in deserts (Reid, 2009).

10.12 Conclusion

River sediments are useful archives with which to reconstruct past hydrologic changes in arid areas. The sediments themselves provide a guide to the type of erosion and

weathering regime in the headwaters. In many parts of the desert world, the late Quaternary alluvial records show a pattern of upward-fining alluvial sequences, indicative of a change from a high-energy and ephemeral or highly seasonal flow regime to a low-energy and less variable flow regime. More useful in certain contexts is the presence of freshwater snail shells within the alluvium (Kröpelin, 1993). Provided they are carefully tested for possible recrystallization, shells within the time range 0–50 ka are amenable to radiocarbon dating. In addition, the stable carbon and oxygen isotopic composition of the shell carbonate can throw some light on the type of environment in which the shells were living (Abell et al., 1996; Abell and Hoelzmann, 2000; Williams et al., 2000; Williams et al., 2006c). Analysis of the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios can clarify the source of the water in which the shells were living and can indicate when and whether certain sub-basins within the main basin were connected to the main drainage basin (Talbot et al., 2000). This applies equally to alluvial clays (Stanley et al., 2003).

However, there are certain limitations involved in using river sediments to reconstruct past climatic changes in deserts, of which the principal one is the inherently fragmentary nature of the alluvial record. This limitation can be offset to some extent by consulting the offshore record (as in the case of the Nile submarine cone), but many desert rivers fail to reach the coast, and in the more arid areas, the alluvium is subject to reworking by wind. A final limitation involved in using fluvial evidence to infer past environmental changes concerns the problems involved in dating alluvial deposits. In large catchments, deposition of alluvial sediments may often be time-transgressive (Vita-Finzi, 1973; Vita-Finzi, 1976), which means that alluvial terraces upstream may differ significantly in age from superficially similar terraces downstream. The solution to this problem is to date the alluvial sequences at a series of sites along the valley using as many independent dating methods as possible. In this context, it is worth remembering that charcoal fragments within the alluvium may have been reworked from previous sediment stores, so the radiocarbon age of the charcoal may be hundreds or even thousands of years older than the time of deposition of the river sediment (Blong and Gillespie, 1978).

In a comprehensive review of process, form and change in arid land rivers, Tooth (2000) illustrates how our understanding of desert river systems is still very limited. In particular, the influence of major flood events on channel form may mean that many such systems are in a state of non-equilibrium. In fact, the concept of equilibrium as applied to rivers in humid areas may not be applicable to desert rivers. A particular gap in studies of desert rivers identified in Tooth's review concerns the almost complete lack of integration between short-term process studies and studies of Quaternary (and older) river history. Tooth concludes with a plea for devising a stronger theoretical basis in regard to dryland river systems in order to improve attempts to manage these systems into the future. This requirement applies

equally to wetlands in deserts. Tooth and McCarthy (2007) reviewed the landforms and sediments diagnostic of wetlands in arid areas and the hydrologic and biogeochemical processes characteristic of such wetlands, and concluded that wetlands in arid areas operate very differently from the more widely studied wetlands in humid regions.

11

Desert lakes

‘What did he say?’
‘He said there was a lake
Somewhere in Ireland on a mountain top.’
‘But a lake’s different. What about the spring?’
‘He never got up high enough to see.’
Robert Frost (1874–1963)
North of Boston: ‘The Mountain’ (1914)

11.1 Introduction

In the geographical heart of the Sahara within the Ténéré Desert of Niger, there is an isolated mountainous ring complex called Adrar Bous (see [Chapter 18](#), [Figure 18.3](#)). If we were to draw an imaginary circle of radius 1,500 km centred on Adrar Bous, the edge of the circle would only just meet the Mediterranean coast to the north and the coast of West Africa to the south. Immediately south of the central granite core of the mountain are the remains of two former lakes, one about 9,000 years old and the other about 7,500–5,500 years old (Williams, 2008). On the floors of both former lakes, there are wind-eroded beds of diatomite and silt. Within the silts, there are shells of freshwater snails indicating that the lakes were permanent bodies of freshwater. Groups of Mesolithic hunter-gatherers lived near the edge of the older of the two lakes until it finally dried out. It then refilled to a lower level, providing water for subsequent bands of Neolithic pastoralists and their herds of short-horned cattle (Williams et al., 1987; Williams, 2008). The occupation sites and fireplaces left by these prehistoric people contain bones of Nile perch, hippo and turtle. At one spot on the edge of the former lake was the partly exposed skeleton of a hippo with a barbed bone harpoon point embedded in its ribcage. Scattered across the Sahara at intervals during the early to mid-Holocene were hundreds of small lakes similar to those at Adrar Bous, offering eloquent witness to a time when the climate was considerably

wetter than today and the regional and local water-tables were close to the surface (Hoelzmann et al., 2004).

The focus of this chapter is on what such lakes can tell us about past climatic changes in areas now arid. We attempt to show how desert lakes have been used to reconstruct past hydrologic and climatic changes, noting instances where purely local factors may outweigh or obscure regional climatic signals. Because this chapter is primarily designed to illustrate some general principles with selected examples and does not aim to be an encyclopaedic compendium of desert lake histories, we will confine our attention to a few specific case studies from different arid and semi-arid areas in Africa, the Near East, Asia and Australia. The desert lakes of the Americas, the Kenya Rift and north-west India are discussed in some detail in the next chapter (Chapter 12) when we come to the long-debated issue of pluvial lakes and their relationship to glacial events.

11.2 Use of lakes to reconstruct past hydrologic changes in deserts

There is an abundant literature dealing with the reconstruction of environmental (including climatic) changes using lake sediments (Haworth and Lund, 1984; Timms, 1992; Last and Smol, 2001a; Last and Smol, 2001b; Smol et al., 2001a; Smol et al., 2001b). The vast majority of these studies concern lakes in humid temperate and sub-arctic latitudes, while studies of desert lakes are far less common (Gilbert, 1890; Gasse, 1975; Degens and Kurtman, 1978; Wasson et al., 1984; Tiercelin, 1986; Williams, 2000; Yang and Williams, 2003; Burrough et al., 2009a; Burrough et al., 2009b; Currey and Sack, 2009a; Currey and Sack, 2009b; Sylvestre, 2009; Yang et al., 2010). In contrast to many big rivers that flow into or through deserts and are often millions of years old, desert lakes tend to be ephemeral features of the landscape with a lifespan in thousands rather than millions of years. They occupy depressions in the landscape that eventually become filled with sediment. Currey and Sack (2009a; 2009b) provide a comprehensive overview of sediment types and depositional processes in desert lakes. Desert lakes may also form as a result of the damming of a drainage channel by a landslide, dune, tufa deposit or a lava flow. Once the lake reaches overflow level, a spillway will be cut into the dam and the lake will soon be drained as the river re-establishes itself. Exceptions to the generalisation that lakes are ephemeral are the deep fault-controlled lake basins of the African Rift Valley, which originated as a result of Miocene and Pliocene tectonic events (Gasse, 1990; Talbot and Williams, 2009), but many of these late Cenozoic lakes dried out well before the onset of the Quaternary period 2.6 million years ago, and much of the earlier evidence has long been faulted, tilted, buried or eroded (Tiercelin, 1981; Tiercelin, 1986; Williams et al., 1986; Cohen et al., 1997; Talbot and Williams, 2009).

Some of the most detailed desert lake records we possess come from the Ethiopian and Afar rift valleys, but here again it is often hard to distinguish between changes



Figure 11.1. Pliocene lake sediments (diatomites) in fault contact with older volcanic rocks, Afar Desert, Ethiopia. (Photo: Françoise Gasse.)

linked to volcanic and tectonic activity (Figure 11.1) and those directly related to climatic change. For example, the change from Pliocene lake sedimentation (Figure 11.2) in the Middle Awash Valley to a regime of river deposition in the Pleistocene (Chapter 10, Figure 10.1) may well reflect tectonic breaching of the dam impounding the lake rather than a change in climate (Williams et al., 1986). Equally, Pliocene Lake Gadeb at 2,300 m elevation in the semi-arid uplands of Ethiopia came into being as a result of lava flows damming the ancestral Webi Shebeli between 2.7 and 2.5 Ma ago (Williams et al., 1979; Eberz et al., 1988). However, the diatomite sequence of former Lake Gadeb reveals three major transgression-regression cycles, none related to tectonic events, culminating in fragmentation of the original lake into a series of shallow pools and swamps (Gasse, 1980), and the pollen preserved within the lake sediments indicates a cooler drier climate around 2.5 Ma (Bonnefille, 1983), consistent with global trends (Williams et al., 1998).

Lakes have been widely used as indicators of previously wetter climates in areas that are now arid. However, the mere presence of lake sediments in the heart of a desert does not necessarily mean that the regional climate was once wetter in that area. In extreme cases, which can be tested experimentally in the laboratory, deflation in the lee of a small rocky hill can lead to exposure of the local water-table and a small lake can form. Furthermore, if a lake receives its water from a distant, well-watered

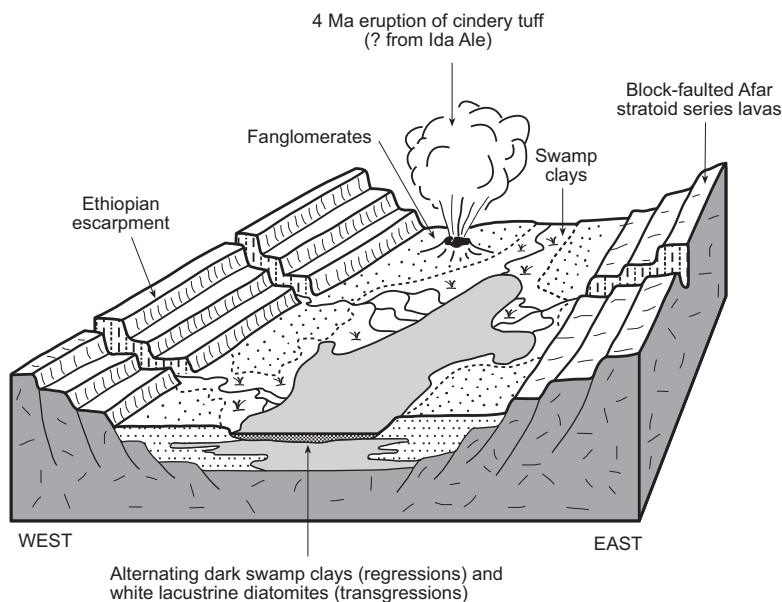


Figure 11.2. Early Pliocene lake deposition in the Middle Awash Valley, southern Afar Rift. (After Williams et al., 1986.)

source, its presence tells us more about conditions in the catchment headwaters than conditions at the site of the lake itself, as hinted at in the quotation from Robert Frost's poem at the start of this chapter. A good example of this is provided by the late Pleistocene Willandra Lakes in semi-arid western New South Wales (Figure 11.3), which were fed by run-off from the Eastern Highlands of Australia via a distributary channel of the Lachlan River (Bowler, 1998; Bowler and Price, 1998). Whether the demise of these lakes was linked to a reduction in run-off from the headwaters of the channel flowing into and through the lakes, or whether run-off was abruptly curtailed as a result of channel avulsion or river capture is still unclear (Williams et al., 1991b; Bowler et al., 2011).

The most certain way to establish previous lake levels is to map the former lake shorelines (Grove and Pullan, 1963; Servant, 1973; Cooke and Verstappen, 1984; Magee, 1998; Magee and Miller, 1998; Drake and Bristow, 2006; Burrough et al., 2009a; Burrough et al., 2009b; Barrows et al., 2014), assuming that they have not been distorted by isostatic (Gilbert, 1890) or other tectonic effects (Flint, 1959a; Flint, 1959b; Burrough et al., 2009b). For a lake to remain stable at any given level, water losses from evaporation, seepage and run-off from the lake must balance water inputs from run-off and precipitation directly onto the lake surface. Thus:

$$A_c P_c k + A_w P_w = A_w E \quad (11.1)$$

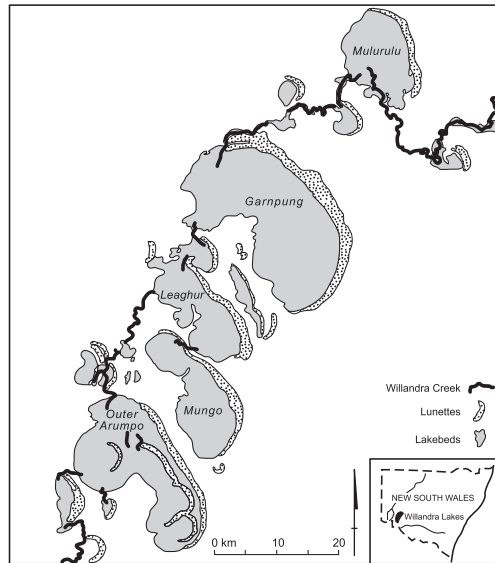


Figure 11.3. The late Pleistocene Willandra Lakes in semi-arid New South Wales, Australia. (After Bowler et al., 2011.)

In this expression, A_c is the catchment area, P_c is the mean annual precipitation over the catchment, k is the run-off coefficient, A_w is the surface area of the lake, P_w is mean annual precipitation over the lake and E is the mean annual surface evaporation from the lake. It follows from this expression that if evaporation is very low and seepage losses are minimal, a lake can sustain a high level even if precipitation is relatively low. Oviatt (2000) offers a variant on this lake water balance equation:

$$\Delta V = A_L(P_L - E_L) + (R - D) + (G_I - G_O) \quad (11.2)$$

In this expression, ΔV is net change in volume of the lake, P_L is precipitation on the lake (expressed as depth), E_L is evaporation from the lake (expressed as depth), A_L is area of lake, R is run-off from the catchment, D is surface discharge from the lake, G_I is groundwater inflows and G_O is groundwater outflows.

11.3 Classification of desert lakes: amplifier and reservoir lakes

Not all desert lakes are sensitive to local climatic fluctuations. Street (1980) discussed the relative importance of climate and more local geological and hydrologic factors influencing lake level fluctuations and drew a useful distinction between ‘reservoir’ lakes and ‘amplifier’ lakes, with the latter being especially sensitive to climatic change. Reservoir lakes are akin to beads on a rosary or pearls on a necklace, and form part of an integrated drainage network, with water flowing in from rivers upstream and out to

rivers downstream. Mega-Lake Chad offers a good example of a reservoir lake; when at its maximum it overflowed via the Benue River into the ocean and received water directly from the Adamawa Mountains of Cameroon as well as the rivers flowing south from Tibesti (Grove and Pullan, 1963; Servant and Servant-Vildary, 1980; Drake and Bristow, 2006). Closed basin lakes operate as amplifier lakes, particularly if the lake catchment area is large relative to actual lake area. In these instances, a very slight change in run-off can cause substantial changes in lake volume. Lake Eyre in central Australia and Lake Ngami in Botswana are excellent examples of amplifier lakes.

Lake status is not, of course, static, and a lake can shift quite rapidly from one state, such as being an ephemeral playa lake, to another, as noted by Bowler (1981) for the desert lakes of inland Australia. In Bowler's (1981) classification of desert lakes in Australia, the ratio of catchment area to lake area is plotted against a climatic function that takes into account precipitation, run-off and evaporation. A change in any one of these three parameters can alter lake status from a saltpan to ephemeral to permanent. The importance of taking into account the ratio of lake basin area to lake area lies in the fact that even a small change in the run-off coefficient can have a substantial effect on lake level if the lake is small relative to its total basin area. For example, if the basin area is 100 km² and the lake area is 1 km², with a mean annual rainfall of 300 mm, a change from 2 per cent to 5 per cent run-off will contribute an additional 9 mm to the lake level each year. If the evaporation from the lake remains very low, this will have an important incremental effect on the rising lake level.

11.4 Desert lake sediments

The sediments flowing into desert lakes are not usually very rich in organic matter, but they may contain appreciable amounts of dissolved silica and carbonate. Depending on the type of river channel flowing into the lake and the distance travelled by the river before reaching the lake, the sediments may consist of sands and fine gravels, as well as silt and clay. The lake margin sediments will tend to be relatively coarse, while those near the centre of the lake will be relatively fine. If the depth of the lake has fluctuated, the sediments will reflect these fluctuations, with coarser material indicative of lake margin facies characterising a *regressive*, or shallowing, lake sequence, and finer sediments overlapping coarser deposits indicating a *transgressive*, or deepening, lake sequence (Currey and Sack, 2009a; Currey and Sack, 2009b; Bowler et al., 2011). The proportions of shallow-water (planktonic) to deep-water (benthic) diatom species can also be used to reconstruct former lake regressions and transgressions (Gasse, 1975). Diatoms are also highly sensitive to changes in water temperature, salinity and alkalinity, and they can therefore be used to reconstruct changes in water chemistry, depth and temperature (Gasse, 1975; Gasse, 1980; Gasse, 1990; Johnson et al., 1990). The fossil mollusc fauna is also a useful indicator of lake history

(Miller and Tevesz, 2001), with permanent water gastropod shells accumulating as beach deposits as the lake level drops (Williams et al., 2003; Williams et al., 2006c). Fossil ostracods are another valuable indicator of water chemistry and temperature (De Deckker and Williams, 1993; Holmes, 2001). If aquatic pollen grains and charophytes are preserved within the lake sediments, they may also be used as indicators of water depth and salinity (Livingstone, 1980). In addition, the calcareous shells of snails and ostracods and the calcareous oogonia of charophytes can all be used for radiocarbon dating (Björck and Wohlfarth, 2001), provided there has been no recrystallization of the original carbonate and that the fossils are no older than about 50 ka. Other methods of dating lake sediments include luminescence techniques [although problems of partial bleaching of quartz grains being dated will need to be resolved (Lian and Huntley, 2001)], electron spin resonance (Blackwell, 2001a) and amino acid racemisation dating (Blackwell, 2001b). Paleomagnetic correlation is also useful but needs independent calibration against other dating methods (King and Peck, 2001).

In some instances, it may be difficult to distinguish between former fluvial and former lacustrine sediments, especially when both consist of horizontally bedded silts and clays rich in aquatic gastropods and ostracods. The finely laminated slackwater silts in the arid valleys of the Sinai and Namib deserts, as well as those in the presently semi-arid Flinders Ranges of South Australia, were all initially interpreted as former lake deposits until detailed topographic surveys established that the gradients of the upper surface of these late Pleistocene fine-grained, valley-fill deposits were parallel to those of the present-day channel floors eroded down to bedrock and were too steep to have been tilted by epeirogenic uplift to that degree in the brief time available (Williams et al., 2001; Williams et al., 2006a; Haberlah et al., 2010a; Haberlah et al., 2010b).

On occasion a river may become a lake. Consider, for example, the lower White Nile, which had an unregulated flood gradient of 1:100,000 prior to the completion of the Jebel Aulia Dam in 1935 near the distal end of that river. As a consequence of its gentle gradient, erosion has been minimal, so that it has a remarkably well-preserved sedimentary flood record. This record spans at least the last 240,000 years and covers two full glacial-interglacial cycles (Williams et al., 2003). During the last interglacial 125,000 years ago, the White Nile formed a lake at an elevation of 386 m (relative to the Alexandria datum) that was more than 650 km long from south to north and up to 80 km wide from east to west (Barrows et al., 2014). This lake was stable at that level long enough for a series of beach ridges several hundred metres wide to develop between rocky headlands located 20–40 km apart. The ridges consist of sandy gravels derived from erosion of the local Precambrian Basement rocks. A second White Nile lake came into being soon after flow resumed from the Ugandan lakes in the headwaters of the White Nile some 14,500 years ago, after a long dry interval spanning the Last Glacial Maximum (21 ± 2 ka) (Williams et al., 2006c).

This lake attained a maximum elevation of 382 m and was up to 20 km wide before it began to recede, leaving behind a scatter of dead aquatic snails. The shells were rapidly buried in several metres of terminal Pleistocene and early Holocene alkaline alluvial clays, and they were therefore preserved as dateable evidence of the 382 m White Nile paleo-lake (Williams, 2012a).

The isotopic record preserved within lake shells, lake sediments and groundwater carbonates also provides an invaluable paleoenvironmental record and can be used to distinguish between lakes fed primarily or solely by local run-off from those fed by groundwater (Cerling et al., 1977; Williams et al., 1987; Abell and Williams, 1989; Gasse and Fontes, 1989; Talbot, 1990; Ayliffe et al., 1996; Ito, 2001; Talbot, 2001). It can also be used to detect major climatic changes within the lake basin. For example, Cerling et al. (1977) deduced that rainfall decreased sharply around Lake Turkana in northern Kenya 1.8–2.0 Ma ago and in the Olduvai Gorge region of semi-arid Tanzania 0.5–0.6 Ma ago. This sudden reduction in precipitation in these two regions was shown by a major increase in the proportion of the heavier isotope of oxygen ($\delta^{18}\text{O}$, see Chapter 7) within pedogenic and groundwater carbonates.

A younger example comes from a series of shallow clay pans in the presently arid region 15 km west of the lower White Nile, in latitude $15^{\circ}22.5'\text{N}$ (Ayliffe et al., 1996). The clay pan sediments contain abundant freshwater gastropod shells dated between 9.9 and 7.6 ka, with most ages (eleven out of fourteen) concentrated at 9.0–8.4 ka (Williams and Jacobsen, 2011), together with some land snails and several semi-aquatic species of snail. Stable carbon and oxygen isotope analysis of the gastropod shells shows that the shell carbonate is highly depleted in ^{13}C and ^{18}O (see Chapter 7). The strongly negative oxygen isotope values show that the region at that time had much lower rates of evaporation than it does today. In addition, the extreme variability in isotopic composition (up to 6–7 per cent PDB) indicates a seasonal rainfall regime with a high degree of variation from year to year. Ayliffe et al. (1996) concluded that there was a stronger south-west monsoon and an associated northward shift of the summer rainfall zone during the time when the clay pans contained water and that this part of the eastern Sahara was both wetter and possibly cooler with lower rates of evaporation around 9.9–7.6 ka. This was also a time when lake levels were high elsewhere in northern Africa, pointing to a regionally wetter climate at that time. Later work has vindicated this claim (Hoelzmann et al., 2004; Williams et al., 2010b; Marriner et al., 2012; Blanchet et al., 2013).

Johnson et al. (1990) provided a concise report on the aims of the International Decade for the East African Lakes (IDEAL) drilling project, together with a summary of the past variations in eight of the main target lakes (Mobutu, Edward, Kivu, Tanganyika, Rukwa, Malawi, Turkana, Victoria) and a brief review of their potential as archives of climatic history. The type of proxy climatic data that the sediments of these (and other) large lakes can provide includes temperature, precipitation, seasonal variability, wind and cloudiness. Temperature can be inferred from grass

cuticle assemblages to a precision of 1°C (Livingstone and Clayton, 1980) and from temperature sensitive pollen taxa, diatom and ostracod taxa, ostracod trace element geochemistry and oxygen isotope analysis of aqueous carbonates (see [Chapter 7](#)). Precipitation cannot be inferred directly but can be estimated from geomorphic evidence of rising or falling lake levels, as well as from diatom and ostracod assemblages, ostracod trace element geochemistry, stable oxygen isotopes and minerals present as indicators of salinity (Fontes et al., 1973; Cerling et al., 1977; Fontes et al., 1985). Sediment laminations and growth layers in oyster shells, mollusca, fish otoliths (Bowler et al., 2011) and stromatolites all offer useful clues to seasonal changes. Wind direction can be deduced from evidence of upwelling recorded in the diatom assemblages and from evidence of anoxic conditions revealed in organic matter, such as sapropel layers (Mee et al., 2007). Cloudiness is hard to assess but may be inferred from light-sensitive diatom assemblages.

11.5 Dating global fluctuations in lake level

One approach to judging the response of lakes in different latitudes to former hydrological changes is that of Street and Grove (1976; 1979), who prepared time series maps of global lake levels during the late Quaternary. Their pioneering work on the paleoclimatic significance of African lakes (Street and Grove, 1976) was later extended to all continents (Street and Grove, 1979). They found that the majority of lakes in the intertropical zone were low or dry during the LGM but were high once more during the early to mid-Holocene. This is a useful first-order approach in documenting desert lake histories, but it depends on very tight chronological control and cannot take account of local forcing factors, such as locally perched water-tables or changes in run-off linked to changes in surface cover. For example, in the southern Negev Desert, Ya'ir (1994) found that run-off is higher today on bare rocky slopes in the more arid southern Negev than it is on the loess-mantled slopes of the central and northern Negev, where infiltration rates are high.

Another factor that needs to be considered when seeking to use lakes as paleoclimatic indicators concerns the influence of extreme events. The scablands of South Dakota bear witness to the sudden discharge from glacial lake Missoula and are of minimal use in interpreting local climate at that time (Baker, 1978; Baker and Bunker, 1985; Teller, 1995). One increasingly exploited archive is the use of evaporite deposits in saline lake basins, because these have excellent potential to record even quite brief climatic fluctuations in the form of solution and precipitation cycles (Wasson et al., 1984; Enzel et al., 1999). A perennial problem concerns the influence of old carbon on the radiocarbon ages and the need to determine the reservoir effects as accurately as possible (see [Chapter 6](#)). One approach, used by Prasad et al. (2009), involved counting annual varves in Late Pleistocene Lake Lisan (the precursor to the modern Dead Sea), which revealed variations in the reservoir age over time.

11.6 Desert lakes of Africa

Scattered across the Sahara are the remains of abundant former lakes ranging in age from Holocene to Pleistocene (Faure et al., 1963; Faure, 1966; Faure, 1969; Williams, 1971; Williams, 1973b; Fontes et al., 1985; Drake et al., 2011). Dating the Holocene and late Pleistocene lakes by radiocarbon analysis is usually straightforward, because the lake sediments often contain mollusc and ostracod shells, biogenic tufas and even charcoal (Williams et al., 1974; Williams et al., 1987; Gasse, 1990; Gasse, 2000a; Gasse, 2000b; Gasse, 2002b). Dating the earlier lake deposits has proven to be more difficult and the results are sometimes hard to interpret, so there are very few detailed studies of older lakes and aquatic ecosystems from this region (Karim, 1968; Williams et al., 1981; Petit-Maire, 1982; Williams, 1984a; De Deckker and Williams, 1993; Wendorf et al., 1993). By way of example, Pleistocene Lake Shati in south-east Libya has provided uranium-series ages considered by Gaven et al. (1981) and by Petit-Maire (1982) to be around 130 ka in age, which would place the lake in Marine Isotope Stage 5 (MIS 5). However, Williams (1984a, p. 440) noted that the uranium-series ages obtained from this site by Gaven (1982) on *Cerastoderma glaucum* shells showing little or no recrystallization (Icole, 1982) fall into four distinct groups. The four oldest samples range in age from 173 to 158 ka, with error terms of up to 20 ka, eleven out of twenty-one samples are dated to 136–132 ka, five are close to 90 ka and the youngest dates to 40 ± 2 ka, so there could equally be four lake phases rather than the single episode inferred by Petit-Maire (1982) and Gaven (1982).

A further example underlines the need for great care when dating lake carbonates. Causse et al. (1988) corrected for the effects of detrital thorium and obtained uranium-series ages of 100–80 ka for lake sediments in the western Sahara considered to belong to the last major wet phase in that area, widely regarded as early Holocene. Szabo et al. (1995) used uranium-series dating of lacustrine carbonates in an effort to obtain ages from the Pleistocene lakes at Bir Sahara and Bir Tarfawi in the Western Desert of Egypt and other former lake sites in the eastern Sahara (Figure 11.4). They identified five discrete lake phases dated to about 320–250, 240–190, 155–120, 90–65 and 10–5 ka. Crombie et al. (1997) subsequently obtained uranium-series ages on travertines from Kurkur Oasis in the Western Desert of Egypt that fell into three broad groups: >260, 220–191 and 160–70 ka. However, all of these ages need to be viewed with considerable caution in light of the earlier experience of Wendorf et al. (1993) in seeking to date the complex of lakes at Bir Tarfawi and Bir Sahara. Wendorf and his colleagues had carefully identified a suite of successive lake deposits associated with Acheulian and Middle Palaeolithic artefacts. They applied a number of different dating methods to these deposits, including luminescence (TL and OSL), uranium-series, amino acid racemisation and electron spin resonance. Out of all of these methods, they found that only the OSL ages yielded stratigraphically consistent results (Wendorf et al., 1993, pp. 552–573). The other methods gave an age range



Figure 11.4. Pleistocene lake marls near Bir Sahara, Western Desert of Egypt.

from 500 to 30 ka, with most of the Bir Tarfawi ages clustered between 180 and 80 ka for what they termed the Grey Lake Phase. They obtained three uranium-series ages for Acheulian sites of 448 ± 47 ka, 542 ± 389 ka and >350 ka (ca. 600 ka), but they rejected all of the ages they obtained for their Acheulian lake sites as either stratigraphically reversed and far too young or with error terms so large as to be meaningless (op. cit., p. 559). However, one important archaeological conclusion did emerge from their work: the inception of the Saharan Middle Palaeolithic is no older than 230 ka (op. cit., p. 358), so the youngest of the Saharan Acheulian industries must be older than 230 ka (see [Chapter 17](#)). The three phases of high lake level documented by Bergner and Trauth (2004) for Lake Naivasha in the Kenya Rift all lie within the interval 175–60 ka and also post-date Acheulian occupation in this region (see [Chapter 17](#)).

A study of lakes can often help explain otherwise obscure changes in river behaviour. The White Nile, for example, was transporting large volumes of sand under conditions of very high energy flow around 30 ka, but its flow had dwindled to a trickle by 20 ka at the height of the Last Glacial Maximum (LGM) (Williams et al., 2010b). The LGM (here defined as the time of maximum global ice volume as deduced from the marine oxygen isotope record, or 21 ± 2 ka: Mix et al., 2001) was a time when the Sahara was even drier than today and desert dunes reached as far south as latitude 12° N (Grove, 1980; Mainguet et al., 1980; Talbot, 1980). However, and more

fundamentally, it was also a time when Lakes Victoria, Albert and Edward in Uganda were dry or at very low levels and no longer flowing into the White Nile (Beuning et al., 1997b; Lærdal et al., 2002; Stager et al., 1986; Stager and Johnson, 2000; Stager et al., 2002; Stager and Johnson, 2008). The resumption of overflow from Lakes Victoria and Albert at 15–14.5 ka (Williams et al., 2006c; Talbot and Williams, 2009) is indicated by the presence of cross-bedded fluvial sands in the lower White Nile Valley, dated by OSL to 13.3 ± 0.9 ka (Williams et al., 2010b), and involved a very different flow regime for the White Nile before the Sudd swamps in southern Sudan had time to become established once more, thereby acting as a gigantic physical filter only allowing the passage downstream of fine silt and clay particles. The abrupt return of the summer monsoon at 14.5 ka was not only seen in overflow from the Ugandan headwaters of the White Nile (Williams et al., 2006c) but also from Lake Tana in the Ethiopian headwaters of the Blue Nile (Lamb et al., 2007; Marshall et al., 2011). This was also a time when the summer monsoon became intensified across tropical Africa (Williams et al., 2006c; Lamb et al., 2007; Gasse et al., 2008; Williams, 2009b; Lézine et al., 2011).

An independent record of climatic fluctuations near the White Nile headwaters is provided by Lake Challa, a crater lake on the eastern flank of Mount Kilimanjaro, which was very high from 10.5 to 8.5 ka (Verschuren et al., 2009), consistent with very wet conditions in the White Nile headwaters at this time and the presence of a large freshwater lake in north-west Sudan at about the same time (Hoelzmann et al., 2000). The four periods of low Holocene lake levels (8.0–6.7 ka, 5.9–4.7 ka, 3.6–3.0 ka, 0.7–0.6 ka) identified by Verschuren et al. (2009) at Lake Challa also coincide with times of low flow in the White Nile. The dry interval starting at 3.6 ka may be coeval with the sharp decrease in rainfall along the southern Dead Sea at around 3.9 ka (Frumkin, 2009), possibly indicating that this arid phase may have been widespread, but until a great deal more evidence is forthcoming, this must remain speculative.

More than fifty years ago, Grove and Pullan (1963) mapped the ancient shorelines of a greatly expanded Lake Chad, later shown to be of late Pleistocene and early to mid-Holocene age (Servant, 1973; Servant and Servant, 1980). Over thirty years later, Armitage et al. (2007) obtained last interglacial OSL ages for the Bama Ridge, which was built during their Lake Mega-Chad phase, and OSL ages of 100–110 ka for Lake Megafezzan in south-west Libya.

Six years after his pioneering work on Lake Chad and more than 4,000 km further south, Grove (1969) mapped the dunes and lake shorelines associated with the distal sector of the Okavango River in the Kalahari Desert of Botswana. Cooke and Verstappen (1984) used air photographs to map what they termed lake ‘Palaeo-Makgadikgadi’ in the depression of that name, today occupied by a series of shallow ephemeral lakes and pans. They estimated that this former lake occupied an area of about 37,000 km² at an elevation of 945 m, and they identified two main lower lake levels at 920 m and 912 m. Within the Makgadikgadi depression, they also mapped

five former deltas of the Boteti River, indicating deposition in a series of former lakes. The Boteti River becomes the Okavango River upstream and arises in the seasonally wet uplands of Angola. The Okavango Delta is one of the largest inland deltas in Africa and is separated from the Makgadikgadi Pans by a major tectonic lineament running from north-east to south-west, called the Kunyere Fault in the south-west and the Thamalakane Fault in the north-east. Cooke and Verstappen (1984) obtained twenty radiocarbon ages from across the Makgadikgadi Basin, sixteen of which were on calcrete (see Chapter 15). They ranged in age from 46 ka to 1.7 ka and cannot be considered very reliable. The challenge of obtaining a reliable chronology was taken up by several groups of workers (Thomas and Shaw, 2002; Huntsman-Mapila et al., 2006; Burrough et al., 2007; Burrough et al., 2009a; Burrough et al., 2009b). Detailed OSL ages are now available for the former high lake strandlines of Lake Ngami (Huntsman-Mapila et al., 2006; Burrough et al., 2007) and the Mababe Depression (Burrough and Thomas, 2008). More than 140 OSL samples show multiple lake full phases for the 'Palaeolake Makgadikgadi', which at its highest recorded level covered an area of 66,000 km². More recent mapping using unusually high-quality helicopter time-domain electromagnetic data suggests an area in excess of 90,000 km² (Podgorski et al., 2013). As the lake level fell, the former lake split into three component basins (Ngami, Mababe, Makgadikgadi), all of which have well-defined shorelines (Burrough et al., 2009a). High shorelines have yielded OSL ages of 8, 17, 27, 39, 64, 92 and 104 ka with relatively small error terms. Beyond that, the error terms increase: 131 ± 11 , 211 ± 16 , 267 ± 27 and 288 ± 25 (Burrough et al., 2009a). Modelling the late Quaternary hydrology of the mega-lake suggests that once the lake attained a threshold size, it developed the capacity to influence both local and regional climate (Burrough et al., 2009b). Huntsman-Mapila et al. (2006) concluded that there was an anti-phase relationship between late Quaternary rainfall in southern Africa and in equatorial Africa, with Botswana dry when the Angolan highlands were wet, much as occurs today. The levels in Lake Ngami were high between 19 ka and 17 ka, at the same time that the central-southern African region showed evidence of increased aridity. Put succinctly, the LGM was arid in Botswana but wet in the Angolan headwaters of the Okavango.

In attempting to reconstruct changes in regional climate, it is important not to rely on the record of any one lake, because local hydrologic influences may sometimes obscure or outweigh the impact of regional climatic fluctuations. Even lakes in close proximity to one another may show quite different response times. For example, Lake Masoko in Tanzania shows a lag of about 1,000 years in maximum inferred effective humidity compared to Lake Malawi, located only 30 km away (Gasse et al., 2008). The late Quaternary history of Lake Masoko during the last 45 ka has been reconstructed from sedimentary and magnetic data (Garcin, 2006; Garcin et al., 2006a; Garcin et al., 2006b; Garcin et al., 2007), pollen (Vincens et al. (2007) and diatom analyses (Barker et al., 2003), all of which indicate driest climatic conditions from 33 to 23 ka,

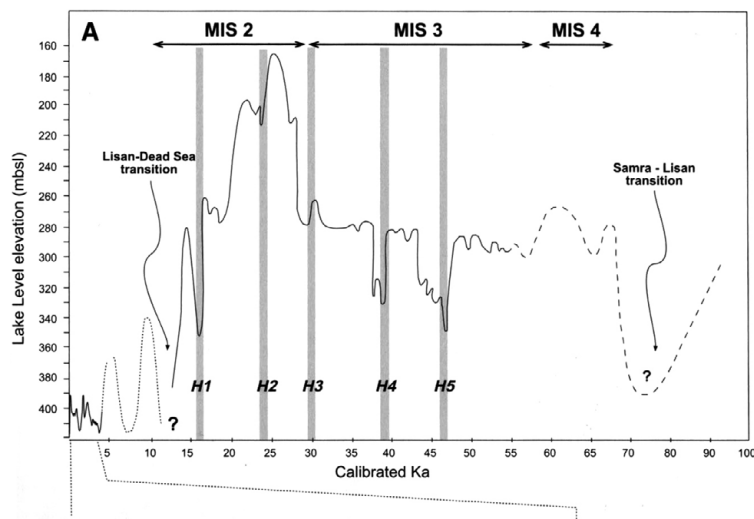


Figure 11.5. Fluctuations in Lake Lisan, the Pleistocene precursor of the Dead Sea. (After Bookman et al., 2006, fig. 4.)

followed by a wet pulse of moderate amplitude at 23–19 ka, coinciding with the LGM. Humidity increased sharply from 16.5 ka onwards, attaining a peak at 13.0–12.5 ka, after which the climate became drier, with a longer and more pronounced dry season. In contrast to the humid LGM at Lake Masoko, Lakes Tanganyika and Rukwa located further north experienced a dry LGM, with a rapid change to wetter conditions by 15 ka (Gasse et al., 2008).

11.7 Desert lakes of the Near East

There are sporadic records of previously high lake levels in the deserts of Syria, Jordan, Arabia and Iran. Of these, the Arabian lakes are the most reliably dated (McClure, 1976; Singhvi et al., 2012), but the overall record is still fragmentary and does not extend very far back, in contrast to the lakes of the African Rift Valleys. One exception to this otherwise patchy record from the Near East is Pleistocene Lake Lisan, the precursor to the present-day Dead Sea, which has received detailed and exemplary geochemical and sedimentological study (see reviews in Enzel et al., 2006) and is also now quite reliably dated (Machlus et al., 2000; Bartov et al., 2002; Enzel et al., 2003; Bartov et al., 2006; Bookman et al., 2006; Enzel et al., 2006; Stein and Goldstein, 2006; Prasad et al., 2009).

This work has demonstrated that Lake Lisan originated around 75 ka from a very low level of about 385 m below sea level (–385 m), rose rapidly to –280 m soon after 70 ka and fluctuated at about that level until around 48 ka (Figure 11.5). The lake level then fell rapidly to –350 m at 46 ka (coeval with Heinrich event H5) and

rose again to -280 m before falling rapidly to -330 m at 39 ka (= H4). The lake rose again to -280 m until 30 ka, when it rose again slightly (= H3). After a slight drop, Lake Lisan rose to its maximum level of -170 m at 25 ka, then fell to -210 m at 24 ka (= H2) before rising slightly to -200 m at 22 ka. The lake level then fell sharply in two stages to -350 m at 16 ka (= H1), rose briefly to -280 m at 14 ka and then dropped abruptly to -380 m, at which time we see a transition from a brackish Late Pleistocene Lake Lisan to the modern hyper-saline Dead Sea (Bookman et al., 2006, fig. 4). Three conclusions may be drawn from the history of Lake Lisan. First, the long interval of relatively constant lake level between about 68 ka and 28 ka indicates strong control over lake level by a sill at this elevation. Second, contrary to expectation, there is no clear correlation between the Heinrich events in the North Atlantic (see Chapter 6) and lake levels, with H1, H4 and H5 coinciding with low levels and H2 and H3 coinciding with high levels. Third, peak levels occur just before the LGM and remained very high throughout the LGM, indicating a pronounced winter rainfall regime at that time. In Turkey, the late Pleistocene Lake Konya was also at its maximum level around 23–17 ka (Roberts et al., 1979). Later work gives a more precise age of 25–20.5 ka for this high lake event (Kuzucuoglu et al., 1999).

Lake Zihor in the southern Negev Desert of Israel is one of the few lakes in this region that date back to the Early Pleistocene. The lake sediments are up to 15 m thick and occupy a tectonic depression that post-dates the Pliocene Arava Formation (Ginat et al., 2003). The central lake facies consists of three beds of white limestone, each about a metre thick, separated by beds of green detrital limestone up to 7 m thick. Between the green and white limestone beds are thin layers of black clay 10–20 cm thick. The white limestone beds contain aquatic mollusca, ostracods and fish-bones consistent with permanent fresh water in a lake estimated to have been 3–5 m deep. The lake dried out on three occasions, and red calcic soils developed on fluvial sediments that interdigitate with the lake deposits. Early Acheulian hand-axes (see Chapter 17) are concentrated around the former lake margins and are typologically similar to those found in the Ubeidiya Formation in the Jordan Valley dated to around 1.4 Ma, as well as to those found at Olduvai Gorge in Tanzania. Early Pleistocene Lake Zihor may therefore have served as a reliable source of fresh water for *Homo erectus* bands moving out of Africa across the Levant and into Asia and would also have attracted small and large mammals. Once the lake dried out permanently, the rivers attracted bands of later *Homo erectus*, as shown by the presence of Evolved Acheulian bifacially worked stone tools on and within the alluvial terraces in this area (Ginat et al., 2003).

11.8 Desert lakes of Asia with special reference to China

We discuss the Holocene lakes of the Thar Desert in north-west India in Chapter 12, so our focus here will be on the desert lakes of north-west China, the Tibetan

Plateau and Mongolia. Both the Aral Sea and the Caspian were considerably larger at intervals during the Quaternary, but these events remain poorly dated. Although not as thoroughly analysed and dated as the desert lakes of Africa, the late Quaternary record of desert lake fluctuations in China is becoming an important paleoclimatic archive.

The deserts of north-western China cover 1.3 million km² and contain sporadic but useful evidence of former hydrologic and climatic changes within this region. For example, in the presently hyper-arid Badain Jaran Desert of north-west China, there are more than 100 permanent lakes nestled among the tallest dunes on earth. This desert is located in the latitude of the westerlies and its southern margin straddles the northern limit of the Asian summer monsoon. Flanking many of these lakes are strandlines denoting former high lake levels. Freshwater mollusc shells are commonly associated with the beach deposits of the lakes, which range in age from early to mid-Holocene on the basis of ¹⁴C and TL dates (Yang and Williams, 2003). Climatic desiccation during the past 4,000 years has led to an increase in salinity within the existing lakes, and water balance models point to a decrease in rainfall from around 200 mm during the early Holocene to around 100 mm today.

The summer monsoon in China actually consists of three independent monsoon systems: the East Asian monsoon coming from the Pacific Ocean, the Indian monsoon coming from the Indian Ocean and the Plateau monsoon coming from the Tibetan Plateau. Consequently, the so-called Holocene climatic optimum, defined as the interval of maximum wetness, was not synchronous throughout this region (An et al., 2000; Yang and Williams, 2003). In fact, there were several episodes of peak summer rains during the early to mid-Holocene, followed by increasing aridity after 5.5 ka and especially after about 4 ka. Five lakes in the Qinghai-Tibetan Plateau region and northern Xinjiang reveal in their oxygen isotope records evidence of an abrupt increase in summer rainfall at 12.5–11 ka that lasted until 8–7 ka, with maximum aridity at all five lake sites from 4.5 to 3.5 ka (Wei and Gasse, 1999). The isotope record was in accord with earlier studies in this region based on sediment, pollen, ostracod and diatom analysis (Fan et al., 1996; Gasse et al., 1996; Van Campo et al., 1996).

Limited evidence shows that certain of the Badain Jaran desert lakes were high around 34 ka, dry after 20 ka and high again by around 13 ka, fluctuating between dry and less dry after that (Yang, 1991; Pachur et al., 1995; Yang, 2001b). In western Mongolia's Valley of the Gobi, the late Pleistocene lakes reached the highest levels, while their Holocene successors were high at around 8.5 ka and were briefly flooded again around 1.5 ka (Lehmkuhl and Lang, 2001). If we allow for minor differences in the time of advance and retreat of the three separate monsoon fronts, the lake histories in this arid region are all reasonably consistent.

Recent work in the Ulan Bui Desert has confirmed these earlier results (Zhao et al., 2012). This desert is located in the arid Alashan region of Inner Mongolia in

north-west China. It lies north of the Helan Shan range and immediately west of the Yellow River. It is bounded to the north-west by the Langshan Mountains and merges north-east into the Hefao Plain. Sporadic outcrops of Holocene lake sediments indicate that this region was once wetter than it is today. Zhao et al. (2012) found evidence of widespread eolian sands until around 8.3 ka, at which time an extensive freshwater lake came into being in the northern sector of the desert, attaining its maximum extent from 7.8 to around 7.1 ka, with a surface elevation of about 1,021–1,026 m. The lake shrank and became segmented into smaller water bodies after 6.5 ka. The modern salt lake at Jilantai (now the centre of a major chemical processing plant) is all that remains today of the once extensive early–middle Holocene desert lake. This work confirms that the climate in this now hyper-arid region was significantly wetter between 7.8 and 7.1 ka than it is today. The area was a sand desert until around 8.3 ka, after which conditions became progressively less arid. The climate became more arid again after about 6.5 ka, and the lake shrank to a small salt lake flanked by episodically active sand dunes.

11.9 Desert lakes of Australia

We saw in Chapters 8 and 9 that based on more than 200 ^{14}C , TL and OSL ages, Bowler (1998) and Bowler and Price (1998) established that eolian dust began to accumulate in the lunettes on the eastern side of Pleistocene Lake Mungo and adjacent lakes from around 35 ka until around 16 ka, with a peak centred around the LGM. Clay dunes and gypseous lunettes were active on the downwind margins of seasonally fluctuating lakes in many parts of south-east and south-west Australia immediately before and between 21 and 19 ka. Major deflation of dry lake floors coincided broadly with the time of extreme aridity centred on the LGM (e.g., Lake Eyre: Magee and Miller, 1998). Gingele and De Deckker (2005) recorded intervals of enhanced eolian dust flux in two cores off the coast of South Australia that span the last 170 ka. During periods of minimum insolation at this latitude, strong northerly winds blew dust from the continental interior, with peaks at about 70–74 ka, 45 ka and 20 ka. These periods accord with times of lake desiccation, dune building and sparse vegetation cover in the centre and south of Australia (Croke et al., 1996).

Although these conclusions are broadly true, more recent work by Bowler et al. (2011) suggests that the earlier paleohydrologic models devised for Lake Mungo and the other Willandra Lakes are in need of some revision. In essence, the Willandra Lakes form a cascading system of what can be considered reservoir lakes fed from Willandra Creek, a tributary of the Lachlan, with its upper catchment in the Eastern Highlands of Australia. Lake Mulurulu was the uppermost lake, which then flowed into Lake Garnpung, which flowed south into Lake Leaghur. The outflow from Lake Leaghur fed into the Outer Arumpo Lake but also fed directly into Lake Mungo (Figure 11.3). Figure 11.6 is a diagrammatic representation of this cascading system.

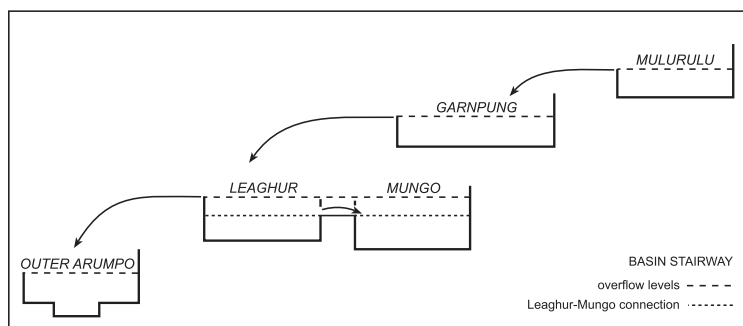


Figure 11.6. The Willandra Lakes viewed as a cascading system. (After Bowler et al., 2011.)

The consequence of this system of basins forming a type of stairway relative to each other was that each individual basin responded quite differently to the fluctuations in discharge from the parent stream, with some basins filling faster and drying out sooner than other basins. As a result, in some basins pelletal clay was being blown from the seasonally exposed floor of increasingly saline lakes, whilst in others the lakes were still deep and fresh and beach gravels were forming under high-energy wave action. This diachronous response to regional changes in climate and river flow is probably true of other reservoir lake systems, such as those in southern Africa fed from the Okavango River.

During the LGM, when the lakes in seasonally wet, tropical northern Australia were mostly dry (English et al., 2001), Lake Eyre in central Australia was totally dry, and its dry bed was being actively lowered by wind erosion (Magee et al., 1995; Croke et al., 1996; Magee and Miller, 1998; Magee et al., 2004). Deflation of desert lake floors during arid intervals may lead to successively lower lake levels after each interval of deflation, giving the impression of progressive desiccation. To eliminate this possible source of confusion, other forms of proxy evidence are needed to test and refute or validate the inferred lake level history. Lake Eyre provides some useful examples. Johnson et al. (1999) analysed the carbon isotopes in fossil emu eggshell from around Lake Eyre in central Australia. They found significant changes in the proportions of C_4 to C_3 grasses over the last 65 ka (see Chapter 7). The data imply that the Australian monsoon was most effective between around 65 and around 45 ka, least effective during the LGM and moderately effective during the Holocene, all of which is entirely consistent with the reconstructed lake levels. Miller et al. (1997) used the temperature-dependent amino acid racemisation reaction (see Chapter 6) in radiocarbon-dated emu eggshells from the continental interior to reconstruct subtropical temperatures at low elevations over the last 45 ka. They concluded that millennial-scale average temperatures were at least 9°C lower between around 45 and 16 ka than they were after 16 ka. There was a sharp change at around 16 ka, followed by rapid warming. These

temperature data suggest that lower temperatures and reduced evaporation may have been a factor in certain lakes being high during LGM times and for the persistence of wetland ecosystems in the arid Flinders Ranges of South Australia (Williams et al., 2001).

Evidence from lakes may also throw some light on other controversial environmental matters. There has been a long and still-unresolved debate as to whether the demise of the megafauna in Australia was the result of human impact (direct or indirect) or of climatic change. The arrival of prehistoric humans in Australia about 45,000 years (45 ka) ago seems to coincide with a wave of faunal extinctions, so on balance it appears that humans were the causal agents. However, the environmental changes at this critical time have always been poorly understood. A recent paper by Cohen et al. (2010b) on late Quaternary aridification and the vanishing of Australia's mega-lakes is a major contribution to this debate, in that it provides, for the first time, unequivocal evidence that desiccation set in shortly after 45,000 years (45 ka) ago. Until then, much of continental Australia was experiencing a very wet climatic phase with vastly expanded lakes in what is now the arid interior. Cohen et al. (2010b) showed that Lake Frome and a series of more northerly lakes were full, and overflowed into a much expanded Lake Eyre at intervals until a final major transgression dated to 50–47 ka, after which lake levels fluctuated and became progressively lower. There were renewed transgressions in Lake Frome late in MIS 3 (around 30 ka), and again at 17, 13, 5 and 1 ka. Using local alluvial evidence and evidence from speleothems in caves located, respectively, in the southern winter and northern summer rainfall zones, they were able to show that southern sources of precipitation contributed to run-off into Lake Frome during the 50, 30 and 17 ka lake transgressions. There was also a tropical contribution to Lake Eyre via the Cooper and Diamantina rivers at 50–47 ka. The 13 ka and younger transgressions appear to represent northerly inputs from tropical sources.

11.10 Conclusion

The early Holocene climates of the tropical northern deserts were generally wetter than they are today, with the highest lake levels occurring around 9 ka. Similar climatic conditions were true of the last interglacial around 125 ka. The desert environments no doubt oscillated between these two extremes, with the interglacials slightly warmer and very much wetter than today and the glacial maxima colder and mostly drier than today. However, not all arid phases coincide with glacial maxima, nor do all humid phases coincide with interglacial times. Some lakes show evidence of humid glacial phases as well as arid glacial phases and humid interglacial phases as well as dry interglacial phases. The evidence from Lake Chad illustrates this well, with high levels at >30 ka to 18 ka, low LGM levels, high levels at 12–9 ka and low levels after 4.5 ka (Servant and Servant-Vildary, 1980). Lake Eyre in Australia was

high at intervals from the last interglacial onwards but dried out completely during the LGM, as did the Willandra Lakes of western New South Wales. However, both of these lake systems receive their water from the Eastern Highlands of Australia and so do not provide direct information about the local climate. The same is true of the Makgadikgadi Lake transgressions in southern Africa, which depended on run-off from the Angolan uplands to the north carried in by the Okavango River and, on occasion, the Zambezi. In China, the desert lakes show differences in the timing of maximum Holocene levels, reflecting the slightly varying influence of the three separate monsoon systems controlling precipitation over the lake basins. In [Chapter 12](#), we examine the evidence for pluvial conditions alleged to be synchronous with glacial maxima, beginning with the pluvial lakes of North and South America.

12

The pluvial debate

And the parched ground shall become a pool, and the thirsty land
springs of water.

Isaiah 35.7

12.1 Introduction

Flood stories are common to many ancient cultures and are epitomised by the account given in the Epic of Gilgamesh, written on clay tablets more than 4,000 ago to describe the life of Gilgamesh, the Sumerian king of the city of Uruk in the lower Euphrates Valley of Mesopotamia, now modern Iraq (Sandars, 1972). Excavations at this and other early Mesopotamian settlements have revealed a sequence of flood deposits and have shown that towns destroyed by floods were later rebuilt and reoccupied, despite being located on flood-prone valley floors. The interest of the Epic stems from the fact that the Sumerians were the first literate inhabitants of Mesopotamia, and their influence, language and writing persisted for many centuries after the demise of their dynasty (Sandars, 1972). It is thus entirely possible that their written account of a great flood seeped into other literary traditions across the wider region, including the Old Testament account in Genesis.

At all events, these biblical narratives had a pervasive influence on later geological thinking, so deposits laid down by melting ice sheets in north-west Europe were often described as ‘diluvial’ and were attributed to the biblical flood. Floods and pluvial events characterised by unusually heavy and prolonged rainfall are thus deeply rooted in the human psyche. By the time that glacial deposits were recognised for what they were, thanks in part to the influence of Louis Agassiz (1807–1873) in Europe and North America, aided by iconoclasts like the vigorous and eccentric Oxford geologist William Buckland (1784–1856), interest in the Great Flood had begun to wane, at least among earth scientists. (Buckland was sufficiently passionate in his belated espousal of the glacial origin of the so-called Drift deposits of central England that

he considered that all geologists who failed to agree with him on this matter – at that time the majority – should be afflicted with eternal itch without benefit of scratching.

The aim of this chapter is to consider how the concept of a pluvial climate first arose and how it has at times led to spurious interpretations, provoking rebuttals and subsequent more rigorous investigations, culminating in our present understanding of the patterns of past climatic change in and around the deserts.

12.2 What is a pluvial?

In common usage, the adjective *pluvial* denotes rainy (from the seventeenth-century Latin term *pluvialis*, itself derived from *pluvia* – the classical Latin word for rain). In geologic usage, the noun *pluvial* often denotes a period of persistent, heavy rainfall and was originally applied to lakes in arid or semi-arid areas that once occupied far larger areas than their present-day remnants, whether they survive today as small freshwater lakes, saline lakes or salt pans.

In *The Encyclopaedic Dictionary of Physical Geography*, Goudie (1985, p. 339) offers this expanded definition and note of caution:

Pluvial: Time of greater moisture availability, caused by increased precipitation and/or reduced evaporation levels. Pluvials caused many lake levels in the arid and seasonally humid tropics to be high at various times in the Pleistocene and early Holocene (hence pluvials may also be called lacustrals), helped to recharge groundwater, and caused river systems to be integrated. Pluvials used to be equated in a simple temporal manner with glacials, but this point of view is no longer acceptable.

Goudie's definition hints at a number of possible problems when attempting to use pluvial lakes to reconstruct past climates in presently dry areas. These issues may be expressed as a series of questions to which we shall seek answers in this chapter. Was the pluvial lake much larger than its modern counterpart because there was far more precipitation in the lake basin at that time? Or was the lake bigger because there was much less evaporation from the lake surface, perhaps related to lower temperatures, at that time? Were the temperatures lower because the lakes were high during glacial intervals, when summer temperatures would presumably have been far colder than they are today? Was the last glacial climate wetter or drier than today? Or was it wet in some regions and dry in others? Finally, how well dated are the lake sediments and shorelines? Our concern here is to offer a brief review of the history of the changing pluvial concept. This overview will serve as a coda to the more detailed discussion in [Chapter 11](#) in which we assessed the climatic insights to be derived from lakes in now arid areas.

Before proceeding further, it will be helpful to define the term *pluvial* more rigorously. Flint (1971, p. 441) provided a comprehensive definition that meets our needs admirably:

In an effort to approach precision, we shall take *pluvial* (noun or adjective) to mean a climatic regimen of sufficient duration to be represented in the physical or organic record, and in which

the precipitation/evaporation ratio results in greater net moisture available for water bodies and organisms than is available in the same area today or in the preceding regimen. Conversely, we shall take *nonpluvial* to mean a climatic regimen in which the precipitation/evaporation ratio is less than that of today, or distinctly less than that of a preceding or following pluvial regimen.

It is clear from Flint's definition that we are dealing with relative changes in precipitation and evaporation that are specific in time and place rather than changes in some arbitrary amount. The distinction is important. An increase (or decrease) in mean annual precipitation of, say, 250 mm (10 inches) will have minimal effect on the vegetation of a tropical rainforest but will have a very significant impact in dry regions if prolonged for decades or centuries.

12.3 Pluvial lakes in North America

Flint (1971, p. 442) mentions that in 1776, Vélez de Escalante discovered shells near Salt Lake in Utah and inferred that a much larger lake once covered the entire area. It was probably the first recognition of what later became known as Lake Bonneville. We have already seen in [Chapter 5](#) that nineteenth-century geologists working in the semi-arid, inter-montane basins in the United States had identified and mapped the shorelines of a series of formerly very large lakes, of which Lake Bonneville (Gilbert, 1890), Lake Lahontan (Russell, 1885) and Searles Lake in California are perhaps the best-studied (Flint, 1971, pp. 446–451; Smith and Street-Perrott, 1983; Lemons et al., 1996; Madsen et al., 2001). The close association between glacial moraines and high-level strandlines recognised by Gilbert (1890) seemed to show that the lakes were high during glacial episodes. Flint (1971, p. 19) noted that a few decades before then, Jamieson (1863) in Edinburgh, Lartet (1865) in Paris and Whitney (1865) in California had all independently arrived at the conclusion that climatic conditions conducive to glaciation in temperate latitudes would have lowered evaporation and lessened aridity to produce higher lake levels in now arid areas, such as the Dead Sea, the Aral Sea, the Caspian Sea and Lake Balkhash in central Asia, and Lop Nor in western China – all of which are now vast saline or brackish lakes. According to Flint (op. cit., p. 20), the term *pluvial* was probably first used by Hull in his 1865 report on the geology of the Dead Sea region.

Flint (1971) summarised what was then known of the Quaternary fluctuations in pluvial lake levels in the present-day deserts of North America. Smith and Street-Perrott (1983) added significant detail to the record twelve years later, while voicing concern over the reliability of many of the radiocarbon ages obtained on samples that had either undergone geochemical change since initial deposition or had absorbed unknown quantities of older carbon. We have discussed some of the pitfalls of radiocarbon dating in [Chapter 6](#), so it will suffice to say here that many of the inconsistencies between lake level histories obtained by different workers and illustrated by Smith and Street-Perrott (1983) arise from problematic age control. What they did show in their

review, however, is that when all dated lake levels are plotted together as histograms showing the number of dates relating to high, low or intermediate lake levels at successive stages in time, then a majority of lakes in presently arid and semi-arid North America had high or intermediate levels during late glacial times. The hypothesis of glacial = pluvial was thus supported for this region on the evidence available at that time.

The unresolved question hinged on the relative importance of all of the factors controlling the water balance in particular lake basins. Earlier studies had begun to tackle this problem. For example, water balance studies by Leopold (1951) and by Antevs (1954) based on late Pleistocene snow-line fluctuations in the mountains of New Mexico suggested that pluvial Lake Estancia was high at that time because of a combination of decreased temperature, decreased evaporation and increased precipitation. Reeves (1965) investigated the pluvial lakes in the Llano Estacado of west Texas and concluded that Pleistocene precipitation was little different from that of today in this region. He concluded that fluctuations in run-off modulated by temperature fluctuations were the primary cause of the high pluvial lake levels in west Texas.

Pleistocene Lake Bonneville is one of the most studied pluvial desert lakes in the world. First investigated in detail by G.K. Gilbert (1890), who identified three well-defined high lake shorelines and evidence of a major overflow channel, as well as deformation of strandlines as a result of isostatic readjustment, at its maximum Lake Bonneville was more than 330 m deep and occupied an area of 51,640 km², with a total water volume approaching 7,500 km³. Gilbert named the high shorelines he identified Bonneville (around 1,565 m elevation), Provo (around 1,470 m) and Stansbury (around 1,350 m). Because of isostatic rebound after loss of lake water, these are best regarded as relative elevations, given that the same strandline will appear at different elevations across the basin. The Provo shoreline started to form after the lake had stabilised at this lower level following the breaching of the alluvial divide, described later in this section.

O'Connor (1993) studied the hydrology of the overflow of Lake Bonneville that took place near Red Rock Pass in Idaho some 14,500 years ago and sought to estimate peak discharge in ten separate reaches along the flood route. Using geomorphic evidence, he estimated that peak discharge amounted to $1.0 \times 10^6 \text{ m}^3 \text{ sec}^{-1}$ at the Lake Bonneville outlet near the Red Rock Pass. Estimated values of stream power below the outlet ranged from $10^1 \text{ watts m}^{-2}$ in ponded reaches to $10^5 \text{ watts m}^{-2}$ in constricted reaches. Stream power (discussed in Chapter 10) was defined by Bagnold (1966) as the rate of energy loss per unit length of stream. Both stream power and sediment transport rate are roughly proportional to stream velocity cubed. Cobbles and boulders carried by the flood ranged from 10 cm to more than 10 m – testimony to the extraordinary magnitude of the Bonneville flood. Once the lake had overflowed and lowered the level of the sill forming its outlet, it was, of course, never able to exceed

that elevation again. We mention this because there is a tendency to consider that older, higher pluvial lake levels necessarily denote previously far wetter conditions in the lake basin. Only in the case of closed lake basins is this line of reasoning valid, and care needs to be taken to establish that apparently closed lake basins were not subject to occasional phases of overflow.

According to Reheis (1999), in the western Great Basin, there seems to be evidence for a solid case that lakes did become progressively smaller from Early to Late Pleistocene. She estimated that the oldest and highest of the pluvial lakes would have needed an increase in effective moisture between 1.2 and 3 times relative to that required by the late Pleistocene pluvial lakes, including Lake Lahontan. Reheis (1999) provisionally correlated the four deep-lake cycles she had identified with marine oxygen isotope stages (MIS) 16, 12, 6 and 2, indicating that all occurred during glacials. This chronology tallies exactly with that obtained independently by Oviatt et al. (1999) in their re-analysis of a core (the 'Burmeister core') in which Eardley et al. (1973) had previously identified seventeen deep lake cycles. The re-analysis only showed four deep-lake cycles, attributed to MIS 16, 12, 6 and 2 (Oviatt et al., 1999). Thirteen of the units previously considered to be lacustrine were found to be marsh or mudflat sediments with associated marsh ostracods and gastropods.

New dates have also been obtained for the Provo shoreline alluded to earlier in this section (Godsey et al., 2005). This shoreline is in fact a composite of multiple shorelines formed during intermittent overflow from Lake Bonneville across the Red Rock sill. It now appears that the shoreline was occupied for longer than envisaged previously, namely from about 14,500 to 12,000 ^{14}C yr BP, dropping rapidly to its present levels by ca. 11,500 ^{14}C yr BP (Godsey et al., 2005). In light of these refinements, the chronology of pluvial Lake Bonneville may be considered a work in progress – which is also true of any lake chronology.

Menking et al. (2004) revisited the question of whether the Last Glacial Maximum (LGM: 21 ± 2 ka) in semi-arid south-west North America was 'wetter or colder'? By colder, they meant dry but with evaporation low enough to sustain high lake levels. They chose Lake Estancia in central New Mexico as a case-study, owing to its well-preserved shorelines and simple hydrology. Using a series of detailed water balance and run-off models, they concluded that in the case of pluvial Lake Estancia, LGM precipitation may have been twice that of today during brief periods of colder, wetter climate and that during those times, annual run-off in the basin may have amounted to 15 per cent of annual rainfall, as opposed to ca. 2.4 per cent in historic times.

Lyle et al. (2012) have provided a stimulating and persuasive contribution to the debate over when and why the Great Basin lakes were full during late Pleistocene times. They compared the timing of wet phases on either side of the Sierra Nevada along roughly north to south transects between 42°N and 32°N . One transect encompassed the coast of California and drew on proxy climatic data from pollen analysis and from alkenone and microfossil estimates of sea surface temperatures adjacent to

the California coast. The other transect ran from New Mexico to Nevada and included recently dated strandlines from Lake Lahontan in Nevada (38–42°N) and Lake Estancia in New Mexico (34.8°N). What they found was contrary to what might be expected if the main source of precipitation in the Great Basin Lakes had come from southward displacement of the winter westerlies during late glacial time. Peak humidity, shown by times of highest lake level, was earlier in Lake Estancia (24.5 to 15.5 ka, with a sharp drop in level at 18–17 ka) and later in Lake Lahontan (17.2 to 14.5 ka, preceded by a sharp drop in lake level between 19.3 and 17.2 ka, after an interval of relatively high lake level between 25 and 20 ka). Furthermore, the central California coast was relatively wet between 12.5 and 4.5 ka, which was approximately 5,000 years *after* the wet interval evident in southern California. They concluded that the dominant sources of precipitation for the Great Basin lakes came from the south, notably from the eastern Pacific and Gulf of Mexico during the summer monsoon season. As more data come to light in the future, this interpretation will no doubt be tested further, but two key points deserve emphasis. First, the onset of peak humidity in the Great Basin lakes was time-transgressive, so that a simple glacial=pluvial equation is no longer tenable. Second, southward displacement of the westerlies can be ruled out as the major cause of late Pleistocene high lake levels in the Great Basin of the western United States.

12.4 Pluvial lakes in South America

The pluvial lakes located at high elevations in the semi-arid Bolivian Altiplano of the central Andes region have been intensively studied by international research teams equipped with the necessary drilling equipment (Sylvestre, 2009). Many of these once freshwater lakes are now saline and are referred to locally as *salars*, or salt pans. Sylvestre et al. (1999) used a combination of ^{14}C and $^{230}\text{Th}/^{234}\text{U}$ ages to determine a lake level chronology for the Uyuni-Coipasa basin. She and her colleagues found that late Pleistocene lake levels began to rise slightly before 16,000 radiocarbon years ago (^{14}C yr BP) (see Chapter 6) and reached maximum levels between 13,000 and 12,000 ^{14}C yr BP. Following a dry spell between about 12,000 and sometime before 9,500 ^{14}C yr BP, the lake rose again to a lower level than in the terminal Pleistocene between about 9,500 and 8,500 ^{14}C yr BP. They obtained good concordance between their ^{14}C and $^{230}\text{Th}/^{234}\text{U}$ ages for the first and highest lake phase but a lack of accord for the second lake phase. They attributed this discrepancy to a delayed response of the groundwater table during the dry phase and used a correction of around 2,000 ^{14}C years for the reservoir effect.

Geyh et al. (1999) sought to establish a reliable ^{14}C chronology for the late glacial/early Holocene humid phase along a high-altitude transect between 18°S and 28°S in the Atacama Desert of northern Chile. They dated a variety of different types of sample, including non-aquatic, carbon-rich sediments, in order to establish the

magnitude of the radiocarbon reservoir effect, which varied between $-1,200$ and a surprisingly large $-10,700$ ^{14}C years, indicating that uncorrected ages would appear far older than they actually were. They found that the humid phase began between $13,000$ and $12,000$ ^{14}C yr BP, with maximum lake levels attained between $10,800$ and $9,200$ ^{14}C yr BP. Their results do not accord with those of Sylvestre et al. (1999) discussed in the previous paragraph in regard to the timing of the wettest phase. One possible explanation for the difference between the two sets of ages bracketing the wettest phase is that the lakes in the two studied areas were influenced by air masses derived from quite different sources operating at different times.

The evidence afforded by desert lakes in South America to help reconstruct past hydrological and environmental changes is a very useful first step towards establishing a history of past climatic changes in presently arid areas. However, the light cast by Geyh et al. (1999) and by Sylvestre et al. (1999) on the magnitude of the radiocarbon reservoir effect and its changes over time causes us to ask whether the frequency analysis of radiocarbon ages for lake level fluctuations in the deserts of North America conducted by Smith and Street-Perrott (1983) may have overestimated the radiocarbon ages of some of the high lake levels in that region. Needless to say, where appropriate steps have been taken to determine this effect over time in a particular lake basin, there should be no cause for concern.

12.5 East African pluvials

J.W. Gregory was among the first to map the geology of the East African Rift Valley. In the course of his fieldwork in southern Kenya in 1893, he found evidence of high lake levels in the form of thick deposits of diatomite on the Kamasian escarpment west of Lake Baringo. Although he considered them to be Miocene in age (Gregory, 1896), L.S.B. Leakey later discovered Early Stone Age Acheulian hand-axes and cleavers associated with these lake deposits, indicating a Pleistocene age (Leakey, 1931). At about the same time, E.J. Wayland, who was then Director of the Geological Survey of Uganda and keenly interested in African archaeology, also noted the presence of previously more extensive lakes and wetlands, and in 1930 he discovered an important prehistoric site on the Kagera River near the Kavirondo Gulf of Lake Victoria. Leakey, who had already embarked on a systematic survey of the Stone Age cultures of Kenya in 1926 (see Leakey, 1931), was quick to recognise that many of his prehistoric sites appeared to be in places that were once much wetter than they are today. Assisted from time to time by Wayland and later by the Swedish geologist Nilsson (1931; 1935; 1940; 1949), Leakey devised a climatic chronology based on supposed pluvials, named after the type localities where they were considered to have occurred. The oldest, or *Kageran*, was considered to be early Pleistocene in age and to be the possible time equivalent of one of the earlier Alpine glaciations, perhaps the Günz. The *Kamasian* pluvial he equated with the Alpine Mindel glaciation; followed by the *Kanjeran* and

Gamblian pluvials, respectively, Mid-Late and Late Pleistocene. The *Makalian* and *Nakuran* wet phases were younger and were believed to be post-Pleistocene in age. Between each pluvial, there was a dry phase, or *interpluvial*. If the temporal equation pluvial = glaciation was indeed correct, then by definition high latitude glaciations were synchronous with low-latitude pluvials.

Almost by sheer force of personality, Leakey ensured that the idea of using pluvials as stratigraphic markers for Africa, at least in a relative sense, was espoused at the First Pan-African Congress on Prehistory, which he organised in Nairobi in 1947. However, Leakey's views did not go unchallenged, although a number of influential scientists did accept the glacial-pluvial model without demur.

Following in the footsteps of Nilsson's earlier work in the highlands of Ethiopia, Büdel carried out a reconnaissance survey of glacial and periglacial limits in the Semien Mountains of Ethiopia in 1953 and concluded that during the last glaciation, the snow-line in this locality had been around 700 m lower than it is today (Büdel, 1954). He also argued that during glacial times, places at lower elevations experienced a pluvial climate and that the tropical deserts in the lowlands were also less arid than they are today during glacial-pluvial phases. The botanist Hugh Scott also visited the Semien Mountains and concluded that '... it appears likely that pluvial periods in the tropics were broadly contemporary with glacial phases elsewhere' (Scott, 1957–1958, p. 11). In this he was particularly influenced by the work of Nilsson (1931; 1940), who referred to the Kanjeran snow-line and the Gamblian snow-line in the Semien Mountains.

However, there were growing numbers of dissenting voices questioning the glacial pluvial concept. For example, Kuls and Semmel (1965) investigated slope mantles in the Godjam Highlands of Ethiopia and concluded that many deposits previously attributed to periglacial solifluction during the so-called 'pluvial period' (*pluvialzeitlicher Solifluktionsvorgänge*) were in fact formed by recent weathering and erosional processes without any need to invoke frost action.

A decade earlier, at the Third Pan-African Congress on Prehistory held in Livingstone (in present-day Zambia) in 1955 and hosted by the archaeologist J. Desmond Clark, concern was expressed at extending a climatic interpretation to describe Quaternary stratigraphic sequences throughout the entire continent of Africa (Clark, 1957; Cole, 1963). The South African geologist Alex du Toit (1947) had earlier expressed grave misgivings in regard to over-enthusiastic cross-continental correlations based on uncritical acceptance of the pluvial chronology, and Wayland had always advocated caution in using the term pluvial. Delegates at the Livingstone Congress accepted that Leakey's pluvial sequence may well have been valid for East Africa but concluded that correlations across Africa needed to be supported by at least two independent lines of evidence: geological, archaeological or paleontological.

At this same conference, the South African geologist H.B.S. Cooke went further and called into question the climatic interpretation of the East African sites. Both

Cooke (1958) and the distinguished American glacial geologist R.F. Flint (1959a; 1959b) provided thorough and dispassionate critiques of the evidence used to infer wetter or drier climates in Africa during the Quaternary, noting that lake fluctuations often reflected volcanic or tectonic influences rather than climate. Flint was careful to draw a distinction between lake fluctuations that *demand*ed a climatic control, as opposed to merely being *consistent* with a wetter (or drier) climate. Other possible non-climatic factors pointed out by both Flint and Cooke were changes in run-off into lake basins caused by changes in river discharge resulting from river capture or the breaching of a natural lake dam by headward erosion. Publication of *Background to Evolution in Africa* (Bishop and Clark, 1967) marked a return to correct stratigraphic procedures. Bishop (1971) provided a history and critique of the East African pluvial concept and was careful to distinguish between interpretation and field observations. He also noted that the Gamblian pluvial was Holocene and not late Pleistocene in age, and categorically rejected the evidence put forward in support of all the earlier pluvials.

Despite these caveats, the notion that glacial climates were wetter than today in East Africa and elsewhere was still well-entrenched (Büdel, 1977). In their comprehensive monograph on *Desert and River in Nubia*, Butzer and Hansen (1968) mapped widespread late Pleistocene gravels in river terraces along the Nile Valley in southern Egypt. They argued that in order to carry such a coarse bed load, the Nile must have had a greater competence and flood discharge than it does today. They concluded from this that climatic conditions were wetter in the Ethiopian headwaters of the Nile during the late Pleistocene, thus supporting the notion of a glacial pluvial climate. It is interesting to note that Fairbridge (1962; 1963) used essentially the same evidence of widespread late Pleistocene alluviation in the Nubian Nile Valley of northern Sudan to argue instead for glacial aridity, on the grounds that the river lacked the competence and flood discharge to carry this material to the sea.

Oddly enough, as we observed in Chapter 10, both groups of workers were partly right. The LGM was indeed more arid in the Ethiopian headwaters of the Blue Nile, but the more seasonal discharge regime led to widespread deposition of sand and gravel in central and northern Sudan and southern Egypt (Adamson et al., 1980). Because the White Nile was cut off from its sources in Uganda, it was no longer able to provide low season flow to the Nile, which became a seasonal rather than a perennial river, so that a great deal of sediment never reached the Mediterranean.

The East African glacial-pluvial model so vigorously espoused by Nilsson (1949) was finally laid to rest at the Pan-African Prehistory and Quaternary Studies Congress held in Addis Ababa in December 1971, at which Karl Butzer and his colleagues presented the first detailed radiocarbon chronology for the Kenyan and Ugandan high lake levels. Their work, published the following year in *Science* (Butzer et al., 1972), demonstrated that the lakes were high during the early to mid-Holocene and were low or dry during the late Pleistocene. Using a simple hydrologic model, they were

able to show that the high early Holocene levels of Lakes Nakuru, Elmenteita and Naivasha in the Kenya Rift would have required a significant increase in precipitation over these lake basins.

A further reason for the final demise of the East African glacial-pluvial chronology was the gradual realisation that the simple fourfold concept of Pleistocene Alpine glaciations embodied in the terms Günz, Mindel, Riss and Würm was greatly oversimplified and that in reality there had been at least ten glacial-interglacial cycles in the last million years (Williams et al., 1998, pp. 23–106).

12.6 Pluvial lakes in Asia

Geologists, explorers and scientific travellers in the 1860s had observed evidence that the vast inland lakes of central Asia, such as the Caspian Sea, the Aral Sea, Lake Balkhash and Lop Nor, had once been even larger in what became known as the pluvial age(s). In the hyper-arid Badain Jaran Desert of Inner Mongolia, north-west China, there are at present more than 100 permanent lakes among the very high dunes of that desert. Many of these lakes are flanked by higher shorelines, some of them dated by ^{14}C and by thermoluminescence (TL) (see [Chapter 6](#)) to early to mid-Holocene in age, when they were far less saline than they are today, and mean annual precipitation was probably at least twice that of today, that is, 200 mm rather than 100 mm, with desiccation setting in about 4,000 years ago (Yang and Williams, 2003). Similar climatic histories are apparent from many sites surrounding this region and were discussed in [Chapter 11](#). All were characterised by relatively arid late glacial climates.

In the Thar Desert of Rajasthan in north-west India, Gurdip Singh analysed the pollen contained in lake core sediments along an east-west transect from wetter to drier. He and his colleagues studied four lakes in particular and concluded that following a long spell of late Pleistocene aridity, these now saline lakes were full and fresh during the early to mid-Holocene, drying out soon after about 4,000 ^{14}C years ago, with the most westerly lakes drying out a few centuries before the lakes in the less arid east of the desert (Singh et al., 1972; Singh et al., 1974; Singh et al., 1990). Singh (1971) speculated that the demise of the Indus Valley Culture resulted from climatic desiccation around 4,000 ^{14}C years BP. This was also about the time of the putative Aryan invasion of India from the north-west – a topic much in dispute. This migration, if indeed it did occur, may itself have been triggered by extreme drought in Persia, Afghanistan and Mesopotamia at that time, for which there is some independent evidence (Cullen et al., 2000; Weiss, 2000).

Later archaeological surveys by V.N. Misra (1983 MS) led him to question Singh's climatic desiccation hypothesis for the abandonment of Mohenjo-Daro situated on the Indus River and Harappa situated on the Ravi River. A number of the Indus Valley sites were located alongside a once active branch of the Ghaggar-Hakra River, fed

in part from the former headwaters region of the Yamuna, Sutlej and Beas rivers Misra suggested that tectonism rather than climatic desiccation caused the demise of these settlements, with tectonic movements disrupting the headwaters of the ancestral Yamuna, which was diverted eastwards to join the Ganga at Allahabad in north-central India, depriving the ancient Ghaggar-Hakra of much of its flow. Attractive though this hypotheses may be, it is not supported by more recent work which has shown that capture of the Yamuna to the east and its loss to the Indus took place between 49 and 10 ka, that is, well before the final collapse of the Harappan centres (Clift et al., 2012). (In Hindu tradition, the ancestral Ghaggar-Hakra was the mythical Sarasvati River, which is reputed to flow underground to the present-day Ganga-Yamuna confluence at Allahabad – today a sacred pilgrimage site for devout Hindus from all over India).

Geochemical analysis by Wasson et al. (1984) of sediments from Lake Didwana in the eastern Thar Desert and sedimentological analysis by Enzel et al. (1999) at Lake Lunkaransar in the now arid western Thar Desert confirmed that the climate was wetter during the early Holocene, indicating a stronger summer monsoon at that time. At present, the south-west summer monsoon provides more than 80 per cent of the annual rainfall in the Thar Desert, and the weaker north-east winter monsoon provides less than 20 per cent (Sikka, 1997; Enzel et al., 1999).

Other evidence of previously wetter climates in the Thar Desert comes from the very extensive deposits of calcrete and the ubiquitous calcareous paleosols in this region, discussed in Chapter 15. In one polygenic dune in the eastern Thar Desert near Lake Didwana, which was excavated to a depth of 18.4 m, there were twelve calcrete layers separated by wind-blown sands, indicating twelve phases of soil development and carbonate precipitation during relatively wet phases, with eleven episodes of dune accretion during relatively dry intervening phases, all within the past 190 ka (Singhvi et al., 2010).

This brief survey suggests that in the deserts of north-west China and north-west India, the last pluvial phase was in the early to mid-Holocene but that the last glacial climate in these areas was in general arid rather than wet. We elaborate on these preliminary conclusions in Chapters 18 to 22 in which we show that the late glacial climates were not uniformly dry.

12.7 Pluvial lakes in Australia

If a lake has remained stable for any length of time and its shorelines have not been subject to tectonic or isostatic displacement, we can define a simple water balance equation (see Chapter 11, Equations 11.1 and 11.2) in which water inputs are in balance with water losses. The water inputs represent direct precipitation onto the lake and run-off into the lake from its total catchment area. The water losses are those from evaporation from the lake surface, seepage from the lake floor and any losses from overflow by the lake. Provided seepage losses are negligible, the level of the lake

may remain high if losses from evaporation are very low, even when precipitation onto the lake and its catchment is relatively low. The example of late Pleistocene Lake George near Canberra, Australia, may be used to illustrate this paradox.

Using the mapped lower limits of periglacial solifluction deposits in the Snowy Mountains of south-east Australia, Galloway (1965b) deduced that the upper limit attained by trees (the ‘tree-line’ or ‘timberline’) during the Last Glacial Maximum had been lowered by at least 975 m. Because the tree-line today coincides with the 10°C isotherm for the hottest month (in this case, January), the difference between the mean January temperature at that elevation today and 10°C is the temperature rise since the LGM, amounting to at least 9°C. He then used the relationship between mean monthly temperature and mean monthly evaporation measured at Canberra to estimate an LGM annual rate of evaporation of 510 mm from Lake George near Canberra. Using the measured ratio between winter precipitation and run-off in a nearby catchment, he estimated that run-off into glacial Lake George would amount to about 104 mm for an annual precipitation of 380 mm, or only slightly more than half the present long-term mean rainfall onto the lake. Galloway concluded that glacial Lake George was up to 30 m deep, despite receiving only half the amount of rainfall received today. This *minevaporal* hypothesis has been widely debated ever since, as has the chronology of the 30 m high lake shoreline. (It now seems more likely, from as yet unpublished new radiometric ages, that the 30 m shoreline may date back to a wetter interval in the late Pleistocene preceding the Last Glacial Maximum).

Galloway (1970; 1983) later extended his reasoning to a re-evaluation of the LGM climate in the south-western United States, discussed in Section 12.3, raising the question of whether the glacial climate was indeed mild and wet, as some palynologists believed (Van Devender and Spaulding, 1979), or cold and dry, the model he considered to be most in accord with the evidence. The issue of whether the pluvial lakes of the south-western United States reflect a wetter or a drier climate is obscured by the problem of estimating the source and amount of run-off into the lakes, some of which represents glacial meltwater.

In semi-arid western New South Wales, the Willandra Lakes – of which Lake Mungo is best known – have yielded a detailed history of high and low lake levels over the past 40 ka, based on both luminescence and ¹⁴C ages (Bowler, 1998; Bowler and Price, 1998). Although all of the ¹⁴C ages are in the process of being fine-tuned at present, it is clear that the LGM was a time of low lake levels and substantial influx of desert dust (*Wüstenquartz*) into the lake floor sediments, which was brought about by regional aridity (Bowler, 1998).

12.8 Glacial aridity in tropical deserts

The alternative hypothesis to glacial pluvial climates in tropical deserts is the concept of glacial aridity. Fairbridge (1965; 1970) was an early and vigorous champion of the hypothesis of glacial aridity and by 1975 there was broad acceptance that late

Pleistocene intertropical aridity had been synchronous in both hemispheres (Williams, 1975; Williams, 1985). It now seems probable that during times of maximum glaciation, the tropical deserts were even drier than they are today and during the interglacial phases, they were somewhat wetter. For example, during the last glacial maximum, many dunes were active well beyond their present limits (Chapter 8), and considerable volumes of desert dust (Chapter 9) were deposited downwind of the desert margins in central Asia and in China, India, Nigeria and Australia (Pye and Tsoar 1990; Williams et al., 1998). These dust mantles are now vegetated and relatively stable.

A similar pattern of glacial aridity is evident also in the Gulf of Aden and the Red Sea (Deuser et al., 1976). The isotopic composition of planktonic foraminifera from deep-sea cores in this region shows that during the last 250 ka, at least, glacial maxima were times of extreme aridity, with increased sea-surface salinity reflecting even higher rates of evaporation than those that prevail there today. We can therefore conclude that in many of the world's hot deserts, the dominant climate during the Last Glacial Maximum (21 ± 2 ka) was drier, windier and colder than it is today, although the summers may still have been very hot. In addition, the desert lakes that had been full and fresh until about 23 ka now dried out or became hyper-saline, previously perennial desert rivers became seasonal and rivers that had once been seasonal became intermittent or ephemeral streams. With glacially lowered sea levels, land areas were greater and so the aridity associated with enhanced continentality was accentuated. Stronger Trade Winds associated with steeper pressure gradients between the equator and the poles caused increased upwelling of cold water close offshore, further accentuating the aridity of coastal deserts. Maximum concentrations of desert dust in deep-sea cores from the equatorial Atlantic Ocean coincide with glacial maxima during the last 0.6 Ma (Parmenter and Folger, 1974; Bowles, 1975) and probably for far longer. Such dust is easily recognised by its high degree of sorting. In a marine sediment core collected off the coast of Mauritania in the western Sahara, deMenocal et al. (2000) found high concentrations of desert dust until the supply was abruptly curtailed at 14.5 ka with the onset of the so-called African Humid Period. [The term is a misnomer because the climate of Africa was not uniformly wet during this period, as Gasse et al. (2008) have demonstrated]. We have enlarged on these topics in Chapters 8 and 9.

In addition to the evidence of enhanced glacial aridity offered by desert dunes and desert dust, many lakes in the intertropical zone were dry during the Last Glacial Maximum, including Lake Bosumtwi, a crater lake in West Africa and thus an excellent paleo-rain gauge (Russell et al., 2003), Lake Tana in the Ethiopian headwaters of the Blue Nile (Lamb et al., 2007), Lake Victoria in the Ugandan headwaters of the White Nile (Johnson et al., 1996; Talbot et al., 2000) and Lake Mungo in arid Australia (Bowler and Price, 1998).

Recent analysis of two Atlantic marine cores, one located immediately west of the Mauritanian Sahara, the other in the North Atlantic within the zone receiving periodic

influxes of ice-rafted debris from melting icebergs during the late Pleistocene, has shown a strong correlation between these events (known as Heinrich events, after their first recognition by Heinrich, 1988) and peaks in dust flux (Jullien et al., 2007). The phases of exceptional Saharan dustiness and aridity probably reflect southward displacement of the Intertropical Convergence Zone (see Chapter 2), presumably reflecting changes in North Atlantic oceanic and atmospheric circulation patterns during Heinrich events.

12.9 Early Holocene pluvial climates in tropical deserts

We have seen that in the late nineteenth and early twentieth centuries, European and American geologists discovered evidence of formerly higher lake levels in the deserts of Asia, North America and Africa. These lakes were at first regarded as pluvial lakes formed during glacial times, which is in general still true of the desert regions of North America. The notion of glacial pluvial climates became solidly entrenched and the so-called pluvial chronology was even used to provide a relative chronology for prehistoric sites in East Africa, with each so-called pluvial being equated with one of the alleged glaciations identified in the European Alps. The far-travelled coarse alluvial deposits of great rivers like the Nile were also interpreted as having originated during glacial pluvial climates.

Although there were dissenting voices, most notably those of Cooke (1958) and Flint (1959a; 1959b), it was not until the high strandlines of the East African Rift lakes (Butzer et al., 1972) and the abundant remnants of former lakes scattered across the Sahara (Faure, 1966; Faure, 1969) were directly dated in the late 1960s and early 1970s that the glacial pluvial concept was finally abandoned for the arid and semi-arid tropics. The African high lake levels were found to be 11,000–9,000 rather than 21,000 calendar years old, that is of early Holocene rather than of Last Glacial Maximum age (21 ± 2 ka). Not only were the early Holocene climates of the tropical deserts wetter than they are today, with highest lake levels around 11–9 ka, but similar climatic conditions were true of the last interglacial around 125 ka. The desert environments no doubt oscillated between these two extremes, with the interglacials slightly warmer and very much wetter than today and the glacial maxima colder and mostly drier. However, not all arid phases in the tropical deserts coincide with glacial maxima, nor do all humid phases coincide with interglacial times. For instance, Lake Chad (Servant, 1973; Servant and Servant, 1980) in the southern Sahara and Lake Abhe (Gasse, 1975) in the Afar Desert of Ethiopia were both very high for at least 10,000 years before 21 ka, when they fell rapidly. They were then intermittently dry (Lake Chad) or constantly dry (Lake Abhe) until 15 ka, rising rapidly thereafter to reach peak levels at 11–9 ka. After about 4.5 ka, both lakes have remained low apart from occasional brief transgressions. The >30 ka to 21 ka phase of high lake levels could be regarded as a humid glacial phase, and the 21 ka to 15 ka regression

could be viewed as an arid glacial phase. Similarly, the early Holocene transgression represents a humid interglacial phase, and the late Holocene interval of low lake levels represents a dry interglacial phase. This simplified fourfold subdivision ignores local hydrological and geomorphic controls over rainfall, run-off, evaporation, seepage losses and groundwater inflow, but it is probably closer to reality than the simple dichotomy between arid glacial and humid interglacial climates.

12.10 Conclusion

From what we have discussed so far, it is evident that the glacial pluvial climates of the North American deserts do not have equivalents in the tropical deserts and semi-deserts of North and East Africa, Asia and Australia. This lack of synchrony prompted Broecker et al. (1998) to argue for what they termed ‘antiphasing’ between rainfall in the Great Basin of North America and the Rift Valley in East Africa. Milly (1999) accepted the primary conclusion of this work but questioned the fourfold decrease in rainfall posited by these workers which reduced Lake Victoria to one-tenth of its present area during the LGM, calculating instead that only a halving was needed. Be that as it may, there now seems to be little dispute that when late Pleistocene lake levels were high in the deserts of the south-west United States, they were low in the lakes of the African Rift Valley. Conversely, when the East African lakes were high during the early Holocene, the lakes were low in the Great Basin of North America.

13

Desert glaciations

The glacier glistens. A distant snow peak scours the mind, but a
snow peak in the tropics draws the heart to a fine shimmering
painful point of joy.

Peter Matthiessen & Eliot Porter
The Tree where Man was Born:
The African Experience (1972)

13.1 Introduction

In Chapters 11 and 12, we reviewed the long history of the debate over whether or not pluvials and glacials were synchronous in the drier regions of the world. We concluded that in North America, there was good evidence that many desert lake levels were high around the time of the Last Glacial Maximum and subsequent deglacial, as were Lake Lisan (the Late Pleistocene precursor of the Dead Sea) and Lake Konya in semi-arid Anatolia in Turkey, both of which reflect more intense winter rainfall at that time.

However, this did not seem to be the case in the tropical northern deserts of Africa and Asia, which showed highest lake levels in the early to mid-Holocene and relatively low glacial maximum lake levels. Indeed, in one recent study of the Quaternary sediments at Dakhla Oasis in the Western Desert of Egypt, Brookes (2010) argued that the record of former temperature maxima at this latitude could be used to date Quaternary pluvial events in this hyper-arid region over the past 800 ka. His reasoning was based on what he called ‘the well-validated premise that temperature cycles in the north African dry belt drive those of precipitation within the seasonally migratory ITCZ’ (op. cit., p. 253), although he did note that there could be ‘a complication arising from delayed surface discharge of pluvially recharged groundwater’ (op. cit., p. 253). That the last interglacial (MIS 5e) was warmer than the Holocene postglacial is shown by current best estimates for eustatic sea level at

that time, which bracket between +5.5 m and +9 m (Dutton and Lambeck, 2012). To achieve the lower limit of +5.5 m implies meltwater contributions from both Greenland and West Antarctica, whereas the upper limit of +9 m requires additional meltwater from East Antarctica (Dutton and Lambeck, 2012). On theoretical grounds, we might expect enhancement of the global water cycle during times of warmer global climate, with higher rates of evaporation from the intertropical oceans and a stronger summer monsoon. Mountain glaciers that were temperature-limited would be small or absent during the last interglacial, while glaciers that were precipitation-limited would grow, provided that temperatures at high elevations were low enough for precipitation to fall as snow.

In the southern deserts of Australia, lakes were almost all low during the Last Glacial Maximum, although some were high at this time, perhaps as a result of lowered evaporation. The Kalahari lakes were high at intervals during the late Pleistocene, as was Lake Eyre in central Australia, but rivers flowing from equatorial and tropical uplands, respectively, fed these two lake systems, so their levels do not necessarily reflect local climatic conditions. In the drier uplands of South America, lakes were high during the terminal Pleistocene and early Holocene but not, on present evidence, during times of maximum regional ice extent.

Two important questions now arise. How reliable is the evidence for former ice activity in the drier parts of the world, and how well-dated is this evidence? Our aim in this chapter is to examine the erosional and depositional records of past glacial activity in arid and semi-arid regions and to discover what they can tell us about former climatic changes. This is especially pertinent in view of current concerns over accelerating glacier retreat in many of the world's mountains. For example, Thompson et al. (2002) examined six ice cores from Kilimanjaro that spanned the last 11.7 ka and identified periods of abrupt climate change at around 8.3, 5.2 and 4 ka, the latter coinciding with a time of major drought. They also found that the small ice caps on this mountain had decreased in area by 80 per cent during the twentieth century, a trend that will lead to their disappearance between 2015 and 2020 if current climatic conditions persist (Thompson et al., 2002; Gasse, 2002a). A more recent UN report (GEAS, August 2012) showed that ice was actively thinning and glaciers were in rapid retreat on all three African mountains with surviving glaciers, namely, Ruwenzori, Kilimanjaro and Mount Kenya. This report also provided a map showing former centres of Pleistocene glaciation in Africa, of which six were in the High Atlas and Jurjura, four were in the Ethiopian Highlands (Figure 13.1), one was in the central Sahara (Mount Atakor in the Hoggar) and four were in East Africa. Absent from this map is Mount Catherine in the Sinai Desert, which also bears evidence of Pleistocene glaciation (Messerli et al., 1980). An obvious question raised by this evidence of Pleistocene glaciers in mountains now devoid of ice is under what climatic conditions were glaciers able to develop in these areas. One further reason why an understanding of past glacial history in dry areas is important arises from the influence



Figure 13.1. Mount Badda, Ethiopia, showing evidence of late Pleistocene glacial erosion.

that glacial and periglacial processes can exert on river discharge and sediment load (see [Chapter 10](#)), as was the case with the Blue Nile (Williams, 2012a) and the rivers of south-east Australia (Williams et al., 2009b).

13.2 Chronology of glacial-interglacial cycles

Before considering the history of desert glaciations, it is necessary to discuss one other point of possible confusion. Mountain glaciers by their very nature tend to erode and destroy the evidence of previous glacial advances. As a result, the early attempts to develop a chronology of Alpine glaciations in Europe depended almost entirely on the evidence left by glacial outwash fans, notably the *Deckenschotter* gravel terraces in the Alpine foreland mapped over many years by Penck and Brückner (1909). Arising from this monumental work was the erection of the once classic fourfold Alpine glacial sequence: Günz, Mindel, Riss and Würm, with older glaciations identified as the Donau and Biber glaciations. Later work revealed that the Alpine *Deckenschotter* were in fact polygenic or composite features, each incorporating the sediments and buried soils of several glacial and interglacial stages (Williams et al., 1993). Away from the Alps, in Europe and North America, a number of regional glacial chronologies were constructed (Flint, 1971, pp. 624–625). These were based on the identification

of interglacial deposits with fossil pollen, snails and beetles indicative of warmer conditions sandwiched between glacial till deposits, some of which contained erratic rocks reflecting long-distance transport by former continental ice caps. However, the fragmentary nature of the terrestrial glacial sedimentary record and the lack of an absolute chronology beyond the limits of radiocarbon dating meant that stratigraphic correlations within and between continents were always open to doubt.

The way out of this impasse came with the pioneering work of Emiliani (1955) and the recognition that the foraminifera preserved in deep-sea sediment cores provided a record of alternating cold and warm sea-surface temperatures. Later work by Shackleton (1967; 1977; 1987) demonstrated that the stable oxygen isotopic composition of the calcareous shells of the foraminifera could be used as a measure of changes in global ice volume and not simply as a measure of changes in ocean temperature or salinity. On this basis, a sequence of glacial-interglacial cycles was identified and numbered from the most recent backwards in time (see Chapters 3 and 6). Thus, the current postglacial is denoted as Marine Isotope Stage 1 (MIS 1) and the immediately preceding glacial maximum as MIS 2. In fact, the Last Glacial Maximum has been defined in two ways. One, used for the sake of clarity and simplicity in this volume, is the time of the most recent minimum sea level coincident with maximum global ice volume, as inferred from the marine isotopic record, and is considered by Mix et al. (2001) to date back to 21 ± 2 ka. The other, based on the revised time of most recent minimum global sea level, constrains the timing of the LGM to between 26.5 and 19.0 ka (Clark et al., 2009). The Mix et al. (2001) age of 21 ± 2 ka falls within the age range of the Clark et al. (2009) estimate of 22.75 ± 3.75 ka, and because the dating of glacial events is seldom very precise, we will continue to use the 21 ± 2 ka age estimated for the LGM.

In this context, it is of interest to compare the timing of maximum advances of mountain glaciers with that of the continental ice caps. A comprehensive review by Gillespie and Molnar (1995) of the timing of mountain glacier advances in widely spaced localities in North and South America, Europe, Asia, Hawai'i, Tasmania and New Zealand revealed that some mountain glaciers advanced much further early in the last glacial cycle, roughly 115,000 to 30,000 years ago, than they did during the LGM, some 20,000 years ago. These authors also concluded that a number of undated glacial landforms, such as moraines, that had been attributed to the LGM could actually be two to four times older. In light of these uncertainties, a reappraisal of the chronology of desert glacial advances is also warranted.

We saw in Chapter 3 that the astronomical factors identified by Milutin Milankovitch (1920; 1930; 1941) have acted as the pacemaker of the glacial cycles and control the length of each cycle (Hays et al., 1976; Imbrie and Imbrie, 1979; Williams et al., 1998). Three astronomically controlled variables influence the amount of solar radiation received from the sun in any given latitude, namely, the distance of the earth from the sun, which reflects changes in the elliptical path of the earth around the sun

(‘orbital eccentricity’), the tilt of the earth’s axis (‘obliquity cycle’), which controls seasonality, and the changing season of the year when the earth is nearest to the sun (‘precessional cycle’). The duration of the orbital eccentricity cycle is 96,600 years and that of the obliquity cycle is 41,000 years. The precessional cycle varies from 16,300 to 25,800 years, and averaged 21,000 years over the past one million years (Williams et al., 1998). In the late Pliocene up until 2.6 Ma ago, the dominant cycles recognised in the marine record were the 23 ka and 19 ka precessional cycles, with the 41 ka obliquity cycles dominating from 2.6 Ma until about 0.7 Ma, after which the 100 ka cycles became dominant (Elderfield et al., 2012). The early glacial cycles were therefore events of relatively low magnitude and high frequency, in contrast to the high-magnitude, low-frequency glacial cycles of the past 700,000 years.

Attempts to determine the precise timing of postglacial mountain glacier retreat and its relationship to global and regional temperature changes have proven controversial. For example, Schaefer et al. (2006) compared ^{10}Be exposure dates for the onset of major retreat after the LGM in mid-latitude mountain glaciers and obtained a mean age of 17.3 ± 0.5 ka for the Southern Hemisphere and of 17.4 ± 0.5 ka for the Northern Hemisphere. From this, they concluded that mid-latitude glacier retreat was synchronous in both hemispheres immediately after the LGM. They observed that the onset of glacier retreat coincided with the start of postglacial *warming* in the Antarctic high-resolution EPICA Dome C ice core record but with a *cooling* trend in the Greenland GISP 2 ice core, where warming did not begin until the onset of the Bølling/Allerød (B/A) interstadial event at 14.7 ka (see Chapter 6). In seeking to explain this anomaly, they postulated that a substantial spread of North Atlantic winter sea ice soon after 17.3 ka would have masked the global summer temperature increase that was thought to have caused the mid-latitude glacier retreat (Schaefer et al., 2006).

Later work by Clark et al. (2009), based on a very large database of 5,704 ^{14}C , ^{10}Be and ^3He ages showed that the Northern Hemisphere ice sheets began to retreat at about the same time (20–19 ka) regardless of size, as did most Northern Hemisphere mountain glaciers. However, they found that the mountain glaciers of Tibet and those of the Southern Hemisphere started to retreat somewhat later (18–16 ka) and the West Antarctic Ice Sheet began retreating later still at around 14.5 ka. They also found that the mountain glaciers in many areas were already at or near their maximum extent by approximately 30 ka, which was roughly synchronous with the time when global ice sheets began to reach their maxima. The time of minimum global sea level (26.5–19 ka) was the interval of 7,500 years in duration when the global ice sheets were in near equilibrium with climate. They attributed the melting of the northern ice sheets to three main forcing factors, namely, an increase in high northern latitude insolation, an increase in atmospheric CO_2 concentration and a rise in tropical Pacific sea-surface temperatures.

An unresolved problem in the study of Quaternary climates is why a long interval of cooling is followed by a relatively short phase of warming, which ends the glaciation. Denton et al. (2010, p. 1652) returned to this question in the context of the Last Glacial Termination and commented that ‘the reduction of [the Northern Hemisphere] continental ice to about its present volume represents one of the largest and most rapid natural climatic changes in Earth’s recent history’. They noted that during the warming intervals associated with rapid melting of the huge Northern Hemisphere ice sheets, sea level rose by 120 m, atmospheric CO₂ increased by 100 parts per million by volume (ppmv) and vast quantities of meltwater entered the North Atlantic, creating stadial (cold) conditions in the Northern Hemisphere and altering the previous oceanic and associated atmospheric circulation patterns. They hypothesised that during each northern stadial, the Southern Hemisphere westerlies shifted poleward, resulting in pulses of oceanic upwelling and warming that in turn caused deglaciation in the Antarctic and Southern Ocean. Future work is needed to test these ideas. The key point here is that many independent factors other than global temperature changes control the waxing and waning of the ice caps and mountain glaciers, so we should expect to find evidence of regional variability.

13.3 Evidence of glaciation

Evidence for former glaciations may conveniently be considered under two headings: the erosional evidence and the depositional evidence. Both sets of evidence are useful and both have certain inherent limitations when it comes to teasing out a climatic signal.

For a glacier to develop, a layer of snow needs to persist all year and become progressively thicker over time. Because cold water can hold more carbon dioxide in solution than an equivalent volume of warm water, melting snow is slightly acidic and will corrode the underlying bedrock. The result is a nivation hollow, a feature that is common above 3,000 m on the northern peaks of Tibesti in the south-central Sahara but that never developed further to form glacial cirques (Messerli et al., 1980). Ice will form when a deep layer of snow becomes compressed by the weight of the overlying snow mantle to form a semi-crystalline mass called *névé*, or *firn*, which is subsequently compressed to form crystalline ice. Ice flows as a result of several distinct processes. Individual crystals of ice under pressure melt at the point of contact, movement occurs under gravity in a generally downslope direction and the glaciers advance slowly down valley. In the case of tropical and temperate glaciers, seasonal meltwater can penetrate down cracks or crevasses in the ice to lubricate the base of the ice at its contact with bedrock. This lubrication can cause quite rapid local movement of the base of the ice column, with the rest of the ice flowing down en masse. In extreme cases, subglacial meltwater can erode into bedrock, and when later exposed



Figure 13.2. Glacially eroded rock-basin lake with moraine dam at outlet, Blue Lake, Snowy Mountains, Australia. (Photo: Frances Williams.)

at the surface, these subglacial channels are often seen to have flowed upslope under extreme hydraulic pressure.

Provided the overall slope of the ice is down valley, basal portions of the glacier can flow upslope across obstacles, scouring and deepening the rock surface behind them to form rock basins. These are especially common at the head of the glacier, where they are termed glacial *cirques*, *cwm*s or *corries*. Once the ice has melted, these hollows often become the sites of rock basin lakes (Figure 13.2). The advancing ice acts somewhat as a bulldozer, eroding any projecting valley spurs and giving rise to the truncated spurs so characteristic of recently glaciated mountain valleys. The typical glacial U-shaped valley cross-section has a dual origin, with postglacial rock avalanches and alluvial fans at the transition between vertical hill slope and relatively flat valley floor. Tongues of diffluent ice can override the valley side at low points on the divide and form glacial breaches, leading to the reversal of local drainage after the ice has gone.

Ice can collect and transport rock fragments that fall from the adjacent mountain cliffs and steep hill slopes. An analogy for what then occurs is a bar of soft soap falling onto sand. The sand becomes embedded in the soap, and washing with it becomes abrasive. Armed with rock fragments, the glacier or ice cap is capable of considerable erosion or glacial abrasion. The result is often a striated or grooved bedrock surface



Figure 13.3. Glacially striated bedrock, Snowy Mountains, Australia. (Photo: Frances Williams.)

(Figure 13.3). The rock fragments in turn become faceted and sometimes striated also. Asymmetric grooves, or chatter marks, aligned perpendicular to the direction of ice movement with the steeper face upslope allow the direction of ice flow to be reconstructed long after the ice has melted. On a related note, asymmetric rocks, plucked on the downslope side and glacially abraded and smoothed on the upslope side, were long ago recognised by French Alpine shepherds and called *roches moutonnées* for their alleged resemblance to the side profiles of distant sheep.

As the glacier advances down the mountain valley, rock fragments derived from avalanches and debris flows will accumulate along the margins of the glacier to form what are called lateral moraines. Moraines are characteristically poorly sorted and often consist of quite coarse and angular rock debris in a matrix of much finer material. The sediment thus formed has a crude bimodal particle size distribution, and so is often termed a *diamicton*. At the snout of the glacier, the debris carried by the ice is deposited in roughly arcuate bands, or terminal moraines, which are deposited as the ice melts and retreats. Dating the outermost of these terminal moraines provides the time of maximum ice advance and, with a greater or lesser time lag, the onset of ice retreat or deglaciation. Erratic boulders (Figure 13.4) may litter the landscape and, in the case of small ice caps, can indicate the source areas from which the ice originated. Once the entire ice mass has melted, the debris previously on and within the ice is let



Figure 13.4. Erratic boulder transported by ice, Snowy Mountains, Australia.

down to form ground moraine or glacial till. The melting of blocks of ice within the glacial till leads to an irregular surface topography, with the resultant deposits often termed hummocky moraine. Small lakes may form within hollows on the surface of the hummocky moraine. As the ice begins to retreat, the resulting meltwater streams will carry and deposit fluvial sediments across the glacial outwash plains. Small lakes are common at the snout of receding glaciers and also become filled with sediment. Dust blown from the glacial outwash plains and deposited downwind is well-known as glacial loess, discussed in [Chapter 9](#).

13.4 Evidence of periglacial processes

On the margins of the high latitude Quaternary ice sheets, such as the Laurentide and Cordilleran ice sheets of North America or the Scandinavian ice sheet of north-western Europe, there was a zone of permanently frozen ground, or permafrost, great swathes of which are still present in Alaska and Siberia as relicts from the last ice age. Along the temperate margins of the permafrost zone and in low latitude mountains, there was (and is) no permafrost. In its stead, there is a zone of periglacial freeze-thaw activity with its own suite of characteristic landforms (Troll, [1944](#); Embleton and King, [1968](#); Davies, [1969](#); Flint, [1971](#); Washburn, [1973](#); Washburn, [1979](#)). These include areas of

stone stripes (Anderson, 1993; Werner and Hallet, 1993) and stone polygons, block fields (*Felsenmeere*) of frost-shattered angular rubble, block streams or rock glaciers and extensive mantles formed by *periglacial solifluction*. Solifluction simply means soil flow, and can result from a variety of processes, such as soil creep, which do not necessarily require frost action. The prefix periglacial denotes soil flow resulting from freeze-thaw activity.

Needle-ice is a very visible aspect of frost action, with small pillars of ice a few millimetres wide and a few centimetres high pushing up small pebbles and frozen soil, which is later distributed downslope as the ice melts during the day. Needle-ice is an effective agent at disrupting the surface plant cover.

Other common periglacial features are *terraces*, or turf-banked terraces, which are caused by frost disrupting the turf or grass cover, with downslope movement of the turf layer so that it overrides the grass cover further down the slope. Vigilance is needed not to confuse these landforms with terraces caused by overgrazing or originating as sheep or goat tracks in upland areas. In some instances, as in the highlands of Botswana, overgrazing can exacerbate the effect of frost action (Hastenrath, 1972; Hastenrath and Wilkinson, 1973). *Thufurs* (Troll, 1944) are small, vegetation-covered mounds about 30 cm high and are characteristic of sites with abundant soil moisture. On gentle slopes in many high mountains, the plants often grow in rough circles up to a metre in diameter.

Where periglacial solifluction is especially active, it tends to smooth out small topographic irregularities in the landscape, filling hollows, and can sometimes disguise the effect of previous glacial erosion. Periglacial processes can also modify glacial moraines, and this is sometimes only evident by examining the micro-fabric (Mills et al., 2009). By removing weathered mantles from irregular bedrock surfaces, periglacial solifluction can in some instances help create *tors*, or isolated pillars of rock. Once again, caution is needed, because tors can form in a variety of different ways and do not require periglacial processes in order to form (Thomas, 1994).

The distribution of these features reflects the temperature prevalent at that particular elevation or latitude. On Mounts Kenya and Kilimanjaro, the lowest occurrence of turf exfoliation from freeze-thaw processes coincides with the 3,500 m contour marking the zone of 0°C temperature minimum, while stone stripes are common on Kilimanjaro above 4,400 m (Hastenrath, 1973).

Both Hastenrath (1974) and Williams et al. (1978) observed a variety of minor periglacial freeze-thaw features above 4,300 m on Ras Dashan in the Semien Highlands of Ethiopia, including stone-banked terraces, stone stripes and polygons, fine-earth polygons, recently frost-shattered boulders and fields of unstable, angular basalt blocks. Neither Büdel (1954) nor Hastenrath (1974) nor Williams et al. (1978) found any evidence of present-day frost shattering or of movement of the resulting angular rubble below about 4,250–4,300 m, which is also the upper limit of tussock grass in the Semien Mountains.

13.5 Inferring past climates from glacial evidence

Many desert uplands now devoid of ice bear traces of previous glacial activity, and where extensive ice caps are present in arid regions such as Patagonia, they show evidence of having been far more extensive in recent times (Murray et al., 2012). For ice to form and persist in desert mountains, the temperature must be sufficiently low in winter for precipitation to occur in the form of snow and sufficiently low in summer for the snow cover to persist and accumulate. There are thus two prerequisites for permanent ice to develop: relatively low winter and summer temperatures and relatively high rates of precipitation in the form of snow. The lower limit of permanent snow is defined as the equilibrium snow-line, or the zone in which summer ablation and winter snow accumulation are in balance, which coincides tolerably well with the lower limit of cirque glaciation (Flint, 1971). The equilibrium line altitude (ELA) is a useful index with which to compare changes in glacial activity over time in any one area and between different regions. In arid mountains, the ELA can vary widely, depending on changes in precipitation. A decrease in snowfall will lead to a sharp increase in the ELA in arid areas. In more humid mountains, changes in temperature become more important influences on the ELA. Osmaston (2005) has provided a useful review of different methods of estimating the ELA and has stressed the need to examine as many individual glaciers as possible on any one mountain, given that aspect, local topography and local climate can cause substantial variation in the ELA in any one region or massif.

Because the snow-line is controlled by two independent variables – precipitation and temperature – we cannot use the lower limits of cirques and glacial moraines to reconstruct either precipitation or temperature. In high mountains in arid regions, where temperatures at high elevations are often low enough for any precipitation to fall as snow, precipitation is often the limiting factor controlling the volume of ice that is likely to accumulate. However, high mountains tend to create their own climate in the form of orographic precipitation. Air masses coming into contact with a high mountain will rise, becoming adiabatically cooler as they do, so that water vapour will reach dew point and condense as rain or ice crystals. The adiabatic lapse rate varies with water vapour content, but cooling rates of $0.65^{\circ}\text{C}/100\text{ m}$ are not uncommon. If we are to obtain the least ambiguous evidence of past climatic change in desert uplands from erosional and depositional glacial landforms, some independent measure of past temperature is necessary.

13.6 Use of periglacial features to reconstruct temperature changes

One reasonably straightforward approach to resolving this dilemma is to use the lower limit of periglacial solifluction deposits to estimate past temperature. For example, if the present lower limit of seasonal freeze-thaw activity on the mountain slopes

coincides with, say, the 4,000 m contour and with a temperature of 10°C in the warmest month, and if the former lower limit of periglacial solifluction deposits was at 3,000 m, then the difference between the temperature at that elevation today and 10°C would give the amount of temperature increase since then. However, long-term temperature measurements are often lacking in desert uplands, so an alternative method involves estimating temperature lowering using plausible lapse rate values.

In the semi-arid Semien Highlands of Ethiopia, frost shattering ('gelifraction') of the bedrock is only active today at elevations above 4,250–4,300 m (Hastenrath, 1974; Williams et al., 1978). Frost-shattered angular rubble has been identified in these mountains at elevations between 3,100 and 3,750 m (Hastenrath, 1974; Williams et al., 1978). Because freezing temperatures would have been needed to shatter bedrock at these lower elevations, the amount of temperature lowering required may be calculated using the present-day mean lapse rate of 0.6°C/100 m measured for the East African highlands, including Ethiopia (Williams et al., 1978). The temperature lowering of between 4°C and 8°C estimated by this method is consistent with estimates of late Pleistocene temperature lowering elsewhere in East Africa that are based on last glacial snow-lines (Flint, 1959b) and on pollen data (van Zinderen Bakker and Coetzee, 1972).

In using periglacial deposits to reconstruct past changes in temperature, it is essential not to confuse such deposits with debris flows and angular colluvium, as noted by Hurni (1982). Periglacial solifluction deposits lack sorting, contain angular clasts with their long axes oriented downslope, have a porous and often fine-textured matrix and are locally derived. They also tend to be coarser towards the surface and finer at depth. In addition, such deposits become thicker and more widespread with increasing elevation (Embleton and King, 1968, p. 513; Davies, 1969, pp. 32–35; Flint, 1971, pp. 275–277; Washburn, 1973, pp. 189–193). In practice, most workers use a combination of both glacial features (erosional and depositional) and periglacial features to reconstruct past changes in the equilibrium snow-line and the changes in the upper tree-line or limit of tussocky grassland, both of which are temperature controlled (Hastenrath, 1972; Hastenrath, 1973; Hastenrath and Wilkinson, 1973; Hastenrath, 1974).

13.7 The glacial record from Africa and the Near East

Messerli et al. (1980) summarised many years of pioneering work mapping glacial landforms in the mountains of North and East Africa. They were able to determine the approximate present-day equilibrium snow-line and the snow-line for the most recent putative maximum glacial advance in the Atlas Mountains of north-west Africa, Tibesti and the Hoggar Massif in the Central Sahara, the Semien and Bale Mountains in Ethiopia, and Mount Kenya, Mount Kilimanjaro and the Ruwenzori Mountains in East Africa. Messerli and Winiger (1992) revisited their earlier work and stressed

the importance of the forest cover and watershed role of the high African mountains from the equator to the Mediterranean in the context of changing climate. Mark and Osmaston (2008) provided a comprehensive and thoughtful review of Quaternary glaciations in Africa, with the main focus on East Africa, which is where the late Henry Osmaston had worked most intensively. They concluded that although existing radiocarbon ages only yield minimum ages for the Last Glacial Maximum advance (21 ± 2 ka), those ages appeared to show that such advances had been synchronous across Africa. However, direct ages of moraines from cosmogenic radionuclides (CRN), in this case ^{36}Cl , are only available for Mounts Kilimanjaro and Kenya, so this conclusion may be premature. Limited chronological evidence from these two mountains suggests that the maximum ice extent since MIS 5 may have occurred close to 30 ka rather than later. In southern Africa, the evidence for late Pleistocene glaciation remains controversial, although recent work in the highlands of eastern Lesotho does provide a plausible case for a limited niche glacier during the LGM (Mills et al., 2009). Once again, the age of the moraines is still not known, so reference to an LGM age must remain hypothetical. This stricture applies equally to many of the periglacial landforms mapped in the drier regions of Africa (Messerli, 1972; Hastenrath, 1972; Hastenrath, 1973; Hastenrath and Wilkinson, 1973). Until they have been dated directly using cosmogenic nuclide dating methods, any attempts to derive information about late Quaternary climates from such evidence must perforce be speculative. However, there is a useful role for speculation in devising testable working hypotheses. In this regard, the attempt by Hastenrath (1972) to compare inferred modern and late Pleistocene snow-lines on a north-south transect across Africa, South America and Australia-New Guinea led him to postulate an equatorward shift of the westerlies during the Last Glacial Maximum, bringing higher rates of precipitation to uplands in temperate latitudes in the Southern Hemisphere.

Besançon et al. (1973) have provided a detailed and comprehensive account of the difficulties involved in efforts to reconstruct former glacial and periglacial processes and associated altitudinal limits in Lebanon. After reviewing all previous work, they concluded that there was a complete lack of accord, and so proceeded to examine every putative glacial cirque in the country. The result of this detailed fieldwork by three highly experienced geomorphologists is sobering: all of the features claimed as glacial, whether cirques or moraines, appear to reflect variations in the limestone bedrock lithology and structure, and owe little to glacial processes. Nivation hollows formed by corrosion beneath snow patches are evident at high elevations where the aspect or wind regime allowed snow to accumulate and persist. Periglacial landforms are likewise poorly developed and could not be used to determine former temperature changes. Their final conclusion is worth noting, for they considered that evidence from pollen analysis, faunal studies and prehistoric archaeology, buttressed by absolute age control, would all be needed to

clarify past climatic changes in this region and so avoid the pitfalls of premature speculation.

Not all glacial landforms are formed as a result of actively moving ice. In some instances, a tongue of diffuent ice may overflow across a col, perhaps creating a glacial breach in the watershed, only to find itself cut off from the parent source of ice. The result is stagnation of the ice mass and progressive melting and downwasting of the ice, resulting in an abundance of meltwater and eventually a deposit of hummocky moraine with frequent *kettle holes* caused by the collapse of ice trapped inside the glacial till. A good example of this process comes from a mountain valley in Anatolia in north-eastern Turkey, in which the glacier had been advancing from before 26 ka until the LGM at 19 ka, only to collapse after 17.7 ± 0.8 ka ('Termination I') (Akçar et al., 2008).

13.8 The glacial record from Asia

Zech (2012, p. 281) commented that 'establishing reliable glacial chronologies in arid regions has been challenging due to the lack of organic material for radiocarbon dating'. Until the relatively recent development of surface exposure dating and optically stimulated luminescence dating (see Chapter 6), the lack of organic material was a major obstacle towards obtaining reliable glacial chronologies from these regions, but this is no longer the case. A growing number of workers are using ^{10}Be surface exposure dating of moraine boulders to date glacial advances in mountainous desert regions, such as the Pamir, Tian Shan and northern Mongolia (Zech et al., 2005; Gillespie et al., 2008; Sanhueza-Pino et al., 2011; Zech, 2012).

The Tian Shan ranges form part of the northern flank of the Tarim Basin, or Taklamakan Desert (see Chapter 8), and they run for about 1,500 km in an approximately east-west direction and rise to more than 7,000 m. The northern and western slopes receive more than 1,000 mm of precipitation a year on average, with moisture derived from the Atlantic and Mediterranean brought by the westerlies during spring and autumn. In winter, the Siberian High blocks the flow of westerly air. The southern and interior slopes of the Tian Shan are very dry and receive less than 300 mm of rain annually, mostly from convectional summer storms. Ice accumulation is therefore limited by precipitation in the south and interior of the ranges and by temperature in the north and west. A comparison of ^{10}Be surface exposure ages from the Tian Shan with those from the Pamir (Zech et al., 2005) indicated moraines dated to 15 ka, 21 ka and >56 ka (MIS 3) in the Tian Shan and extensive last interglacial moraines (MIS 5) in the Pamir (Zech, 2012). The Pamir Mountains lie south of the Tian Shan and form the western flank of the Tarim Basin. A plausible interpretation of these differences is that the Tian Shan MIS 3 glaciers reflect an increase in westerly precipitation, while the more southerly Pamir received increased monsoonal rainfall during MIS 5. The

ice was more extensive in the Tian Shan during MIS 4 than during MIS 2, indicating increasingly arid conditions in Central Asia during the last glacial cycle (Zech, 2012). This conclusion accords with the evidence of increasing aridity in the Pamir inferred by Zech et al. (2005) to account for the progressive reduction in late Pleistocene glaciation over time.

An interesting new approach to checking glacial chronologies involves dating boulders from very large landslides. Sanhueza-Pino et al. (2011) used this approach in three formerly glaciated valleys in the Kyrgyz Tian Shan. The ages of the landslides were obtained using ^{10}Be exposure dating methods, and care was taken to establish that the landslides showed no signs of having been exposed to glacial erosion since their accumulation. The ages obtained for the three landslides were 67–63 ka, 15–11 ka and 8–6 ka, and they provide a minimum age for glacier advances in their respective valleys. They also mark the maximum extent of the ice within those valleys. The MIS 4 age for one of the glacial advances again confirmed that ‘glaciations in the Tien Shan are distinctly asynchronous with regard to glaciations in Europe and North America’ (op. cit., p. 303). They also concluded that previous estimates for minimum equilibrium snow-line altitudes for the northern Tian Shan needed to be revised upwards by about 400 m.

Gillespie et al. (2008) mapped three main late Pleistocene glacial advances in the Darhad Basin in northern Mongolia and were able to date two of them reasonably accurately. They used a combination of ^{14}C dating, cosmic ray exposure dating (^{10}Be) and luminescence (IRSL) dating methods in order to obtain a reliable chronology of times of maximum local ice advance and times of ice retreat. The two most recent advances had ages of around 53–35 ka (MIS 3) and around 19–17 ka (MIS 2), synchronous with advances of similar extent across northern Mongolia but different from glacial advances in Siberia and western Central Asia. An older and more extensive glaciation possibly dates back to MIS 6, but there was remarkably little difference in the equilibrium line altitude (ELA) for all three glaciations. In contrast to the Tian Shan, where enhanced aridity during the LGM confined glaciers to high elevations at that time, glaciers advanced to relatively low elevations during MIS 2 in the Darhad Basin of northern Mongolia, indicating less arid conditions in that region at that time.

The Tibetan Plateau is the largest high plateau on earth, and covers an area of 2.6 million km^2 with a mean elevation of 4,600 m. Opinions about the extent of the ice cover on the plateau during the LGM differ radically, with most workers arguing for limited and spatially isolated glaciations (Owen, 2009), but Kuhle (2001; 2002) claimed that an ice cap covered most of the plateau. An independent test of these conflicting scenarios is provided by a detailed ecological analysis of the Alpine Steppes of the Tibetan highlands (Miehe et al., 2011). The high proportion of endemic species in this biome argue for prolonged stability during the LGM, and

both pollen and $\delta^{18}\text{O}$ proxy data support the view that the Alpine Steppe persisted during the LGM when temperatures at these elevations were probably about 3–4°C lower than today. Although some decrease in LGM precipitation seems possible, conditions of extreme LGM aridity on the Tibetan Plateau are ruled out by the inferred persistence of the Alpine Steppes in this area during the late Pleistocene (Miehe et al., 2011).

13.9 The glacial record from the Americas

The Andes in South America and the Rockies in North America are unlike any other major mountain ranges on earth in that they run in a more or less direct line from north to south, parallel to the meridian. (The Eastern Highlands of Australia also run from north to south, but with a maximum elevation of 2,228 m, they scarcely qualify as a major mountain range.) The Andes extend for some 7,000 km from the zone of westerly precipitation in the far south to the zone of easterly precipitation controlled by the seasonal migration of the ITCZ in the centre and north (see Chapter 21). As a result, the rain-shadow effect enhances aridity in the Patagonian Desert, located east of the Andes in the south, and in the Atacama and Peruvian coastal deserts, located west of the Andes in the centre and north of the continent. We might therefore expect that glacial activity in the Andes will reflect summer easterly precipitation in the tropical north and winter westerly precipitation in the far south. In addition, the location of the rain-shadow zones will vary over time in accordance with changes in the location and persistence of the southward movement of the ITCZ in summer and the northward movement of westerly air masses in winter. Such changes were not necessarily synchronous.

The Rockies extend for about 5,000 km and are located further from the equator than the equatorial northern Andes. They also display a slightly different pattern of air mass movement, with a much stronger influence from the westerlies, bringing winter precipitation and summer aridity to California but year round rain to Oregon and Washington. (The Cascades and the Sierra Nevada are two smaller, separate ranges that are aligned roughly parallel to the main Rockies.) Another important difference between North and South America concerns the depth and extent of ice caps during times of maximum glaciation. During the Last Glacial Maximum, much of North America was hidden beneath a vast layer of ice up to 3 km thick – the Laurentide ice sheet. This ice sheet (and, no doubt, its predecessors) was sufficiently high that it seems to have split and diverted the high-level jet streams from their customary interglacial (and present-day) mean positions over the continent, thereby altering the circulation patterns at lower levels in the atmosphere (Kutzbach and Wright, 1985). However, whether this led to a significant southward displacement of the westerlies during the LGM now seems very unlikely (Lyle et al., 2012). Both the Andes and the

Rockies are sufficiently high to have perennial snow and ice. The Andes have a mean elevation of about 4,000 m and rise to 6,962 m, while the Rockies are much lower and are mostly below an elevation of 4,000 m.

One much-debated question is whether or not glacial advances and retreats during the past 20,000 years have been synchronous or out of phase in both hemispheres. As far as major ice caps are concerned, the issue seems to be resolved, with ice sheet expansion over Greenland, North America and north-west Europe coinciding with expansion of the East and West Antarctic ice sheets. However, mountain glaciers will often show a far more variable response owing to differences in size, shape and aspect, as well as local controls over precipitation.

The Holocene glaciers in the Peruvian Andes have several sets of moraines, of which the two most prominent have provided high-precision cosmogenic ^{10}Be surface exposure ages. The older moraines date between 10 and 8 ka and are broadly synchronous with early Holocene glacial moraines in the southern Andes, Norway and the Austrian Alps but are out of phase with ^{10}Be -dated Holocene glacial maxima in New Zealand (Licciardi et al., 2009). The younger set of moraines date to the latter part of the Little Ice Age, dated in Europe to between around 1300 and 1860 AD (Lamb, 1977; Grove, 1988). Licciardi et al. (2009) proposed that their results suggested a climatic link between the North Atlantic and the Peruvian Andes, with cold conditions in the north promoting glacier expansion in Europe and a concomitant southward shift of the ITCZ bringing increased snowfall to the Peruvian Andes and glacier expansion. Jean Grove (2004) had earlier demonstrated that there were multiple glacial advances in both hemispheres during the Holocene, so future work needs to focus on dating a wider array of localities in order to provide a more comprehensive and robust set of data, which would allow more rigorous modelling of climatic linkages between hemispheres.

If we go slightly further back in time to the terminal Pleistocene, glaciers in the arid Andes of south-west Peru show signs of a readvance or at least a prolonged stillstand, with moraines located about midway between present and LGM limits dated by cosmogenic ^3He to 12.8 ± 0.7 ka (Bromley et al., 2011).

Ramage et al. (2005) have evaluated the pitfalls involved in using different methods to determine the ELA in a presently ice-free part of the tropical Andes of central Peru. They concluded that best estimates for the LGM amount to a lowering of 220–550 m to between 4,250 and 4,570 m elevation, indicating a modest temperature decrease of around $2.5 \pm 1^\circ\text{C}$. They also noted that the ELA lowering inferred for the LGM was little different from that of the most extensive glaciations in these valleys, which date back to >65 ka. This would suggest that the relative influence of temperature and precipitation on snow accumulation was not the same at these two times. It is interesting to note that the estimates for ELA lowering in the Peruvian Andes during the LGM differ quite markedly from those north of the equator in the Venezuelan Andes (Stansell et al., 2006). Here the ELA levels were around 1,420 to 850 m

lower than present, indicating that temperatures may have been $8.8 \pm 2^\circ\text{C}$ cooler than today.

^{10}Be exposure ages for boulders on the crests of a succession of moraines laid down by former glaciers in the Rio Guanaco Valley of southern Patagonia at latitude 50°S show that the last local glacial maximum had ended by 19.7 ± 1.1 ka (Murray et al., 2012). Rapid glacier retreat was underway by 18.9 ± 0.4 ka, and more than half of the upper valley ice retreat was accomplished by 17.0 ± 0.3 ka. An important conclusion arising from this work is that glacier retreat in southern Patagonia was linked to high-latitude warming in the Southern Hemisphere, associated with changes in ocean circulation initiated by the retreat of the large Northern Hemisphere ice sheets at around 19 ka, thereby generating a ‘bipolar seesaw’ response (Blunier and Brook, 2001; Steig, 2006).

García et al. (2012) obtained thirty-eight ^{10}Be exposure ages from moraines of outlet glaciers in the South Patagonian Ice Field at 51°S , which showed that they advanced during the Antarctic cold reversal (14.6–12.8 ka) and were in rapid retreat by 12.5 ka, consistent with temperature changes in Antarctica and the Southern Ocean at that time. The primary source of precipitation was from westerly air masses, which reached much further north during the LGM, when the Subtropical Front reached to about 40°S , and fluctuated thereafter in response to changes in the Sub-Antarctic Front and the Polar Front (García et al., 2012).

Turning now to the drier regions of North America, it is reasonably certain that nowhere else in the desert world is the evidence of former glacial activity more thoroughly studied. The volume edited by Porter (1983) provided a comprehensive overview of Late Wisconsin mountain glaciation in the western United States during the late Pleistocene, including the LGM. At the time, chronologies were based primarily on radiocarbon dating, so glacial deposits were in general only dated indirectly. Increasing use of cosmogenic nuclide exposure dating during the past decade or so meant that, for the first time, glacial moraines could be directly dated, although here, too, some degree of caution is necessary, as outlined in Chapter 6.

Present evidence shows that mountain glaciers advanced close to their maximum limits during the LGM, but many individual glaciers reflect the influence of local topographic factors and do not respond solely to regional climatic controls. Young et al. (2011) obtained ^{10}Be surface exposure ages from moraines, bedrock and river terrace gravels relating to the late Pleistocene Pinedale glaciation from three adjacent valleys in the upper Arkansas River Basin. They found that times of maximum glacier expansion were not synchronous and ranged in age from 22 to 16 ka. Ice retreat, in contrast, began at the same time, at 16–15 ka. They concluded from their study of Pinedale glacial moraines at widely different sites in the western United States that ‘glacier fluctuations are ultimately driven by climate change, but the exact position of a glacier terminus is filtered by nonclimatic factors intrinsic to each glacier valley system’ (op. cit., p. 173).

13.10 The glacial record from Australia

The semi-arid Snowy Mountains in south-east Australia were glaciated on a number of occasions during the late Pleistocene, but the area covered in ice during the most recent glacial advances did not exceed about 50 km² (Galloway, 1963). Barrows et al. (2001; 2002; 2004) obtained ¹⁰Be and ³⁶Cl exposure ages of 55–65, 32 ± 2.5 , 19.1 ± 1.6 and 16.8 ± 1.4 ka for the youngest glacial advances in the Snowy Mountains and ages between 23 and 16 ka, with a weighted average age of 21.9 ± 0.5 ka, for the much more extensive periglacial deposits in this region. The 19 ka, or Blue Lake Advance, was also synchronous with the LGM ice cap advance in Tasmania (Colhoun et al., 2010; Colhoun and Barrows, 2011), which was also the time of maximum periglacial activity. The 55–65 ka glaciation was more extensive than later ones, perhaps reflecting an increase in aridity and decrease in snowfall during the late Pleistocene.

There is an interesting contrast between late Pleistocene glacial and periglacial activity in Tasmania, which, on a very small scale, mirrors events across North America and Eurasia for similar climatic reasons. In the humid west of Tasmania, the LGM glaciers reached to within 300 m of the present sea level and 420 m of the LGM sea level. In the drier east of Tasmania, the LGM snow-line was much higher and dunes were active close to the coast. Colhoun (2000) estimated that the LGM orographic snow-line in Tasmania was 690–1,000 m lower than it is today, with an average of 830 m for the ice cap on the central plateau.

Galloway (1965b) had earlier mapped the lower limits of periglacial solifluction deposits in mainland south-east Australia. From this, he deduced that the upper limit attained by trees (the ‘tree-line’ or ‘timberline’) during the LGM had been lowered by at least 975 m. He concluded that because the tree-line today roughly coincides with the 10°C isotherm for the hottest month (in this case, January), the difference between the mean January temperature at that elevation today and 10°C is the temperature rise since the LGM, amounting to at least 9°C. Galloway (1965b) concluded that the lower limit of LGM periglacial solifluction, which primarily reflects temperature control, had been lowered by around 600 m in both eastern and western Tasmania. As already noted, this limit is broadly equivalent to the 10°C isotherm for the warmest month. Depending on the lapse rate used, it indicates an LGM temperature lowering during the warmest month of about 5°C. This figure is lower than that estimated for the Snowy Mountains, where the orographic snow-line was 600–700 m lower, the lower limit of periglacial solifluction was at least 975 m lower and the temperature in the warmest month was at least 9°C cooler (Galloway, 1965b).

Deglaciation followed the time of coldest sea surface temperatures in the west Pacific, and there was no evidence of any glacier readvance during the Antarctic Cold Reversal (around 14–13 ka) or the Younger Dryas (around 13–11.5 ka) (Barrows et al., 2001). Blue Lake (Figure 13.2) in the Snowy Mountains was free of ice by 15.8 ka,

suggesting that deglaciation was rapid. Barrows et al. (2001) noted that the glacial advances in the Snowy Mountains of Australia were more or less synchronous with glacial advances in Tasmania, New Zealand and South America, indicating a common response to cooler late Pleistocene climates in the Southern Hemisphere.

13.11 The glacial record from Antarctica

It is easy to forget that Antarctica is the driest continent on earth – it is simply too cold for much precipitation. Australia is the second driest continent, and the contrasts between these two continents could not be greater. On the surface of the Dry Valleys of Antarctica, wind-abraded pebbles, or ventifacts, are common; they are identical in form and origin to the better-known ventifacts of the Atacama, Namib, Gobi and Sahara deserts. However, the Antarctic sedimentary record of past climatic events is patchy and poorly dated, in contrast to the long and unrivalled record from ice cores drilled to several kilometres depth into the central ice cap of East Antarctica, such as the cores collected from near the scientific base at Vostok and jointly studied by Russian and French glaciologists. These ice cores provide a million-year record of past fluctuations in the global atmospheric concentration of methane and carbon dioxide ($p\text{CO}_2$), as well as the more regional temperature changes inferred from fluctuations in deuterium, the heavy isotope of hydrogen. They also provide a detailed record of past fluctuations in the atmospheric dust flux, discussed in [Chapter 9](#).

What these records clearly indicate is that during times of minimum temperature (i.e., glacial maxima), the $p\text{CO}_2$ levels hovered around 160–180 parts per million by volume (ppmv), rising to 260–280 ppmv during the warmest times (i.e., interglacials). These background records are sobering because they reveal that since the onset of the Industrial Revolution around 1750 AD, the $p\text{CO}_2$ levels have been increasing at an accelerating rate, and by April 2013 they exceeded 400 ppmv. Carbon dioxide is a potent greenhouse gas, meaning that it has the capacity to allow the passage of short wave solar radiation through to the earth's surface but will absorb a certain proportion of the outgoing long wave, or terrestrial, radiation, thereby causing slow but inexorable warming of the lower atmosphere (see [Chapter 25](#)).

The atmospheric concentration of two other greenhouse gases has also been increasing exponentially during the last 200 years, namely, methane and nitrous oxide, adding to the absorption of outgoing infrared radiation and hence enhancing the warming of the lower atmosphere. We deal with some of the climatic, ecological and social repercussions of these changes in atmospheric chemistry in [Chapters 25 and 26](#).

Another long-debated question was whether or not the Antarctic and Greenland ice core records were in or out of phase. The issue was resolved in 2006 when an ice core was obtained from Dronning Maud Land in Antarctica with a resolution comparable to the Greenland ice core records (EPICA Community Members, 2006). The results showed a temporal correspondence between Antarctic warm events and Greenland

cold events, consistent with the ‘bipolar seesaw’ hypothesis (Blunier and Brook, 2001; Steig, 2006). The authors concluded that both sets of changes (in Greenland and in Antarctica) reflected a reduction in the overturning of cold surface waters flowing south from the North Atlantic (EPICA Community Members, 2006).

13.12 Conclusion

For mountain glaciers and ice caps to form, two conditions are necessary. The temperature needs to be sufficiently low for precipitation to occur in the form of snow, and there needs to be sufficient precipitation for the winter snow to persist throughout the following summer and for this process to continue until the snow at depth is converted to ice. The elevation at which net ice accumulation within a mountain glacier and net loss of ice through melting are in long-term balance is termed the equilibrium snow-line, or the equilibrium line altitude (ELA). In the drier regions of the world, such as the Tian Shan ranges in Asia, the ELA is highly sensitive to even small changes in precipitation and will be at much higher elevations during dry climatic intervals than during wetter phases.

There is growing evidence from glacial moraines, directly dated using cosmogenic nuclide exposure dating methods, that in many arid and semi-arid areas, the Last Glacial Maximum at about 21 ± 2 ka was not the time of maximum glacier advance but was preceded by more extensive glaciation earlier in the Late Pleistocene or even before then, suggesting an increase in aridity during the Late Pleistocene and perhaps also during successive glacial cycles. This inference remains a working hypothesis, but is supported by the evidence from desert lakes outlined in [Chapter 11](#).

Because temperature is such a critical factor controlling glacier advance and retreat, too narrow a focus on purely local climatic influences can be unhelpful. Evidence from the central Andes in South America suggests that at least during the Holocene, the climatic linkages responsible for the waxing and waning of mountain glaciers may have been more closely linked to events in the North Atlantic than they were to more southerly influences. In southern Patagonia, there is also persuasive evidence that the rapid ice retreat there, which began at 19 ka, was associated with Southern Hemisphere warming triggered by Northern Hemisphere forcing operating via the ‘bipolar seesaw’ phenomenon. This phenomenon, discussed in [Section 13.11](#), simply means that when temperatures in Greenland and the North Atlantic were cold, temperatures in the Antarctic and Southern Ocean were relatively warm, and conversely. A further example of this effect is the terminal Pleistocene ice advance in southern Patagonia which reached its maximum extent by 14.2 ka, during the Antarctic cold reversal (14.6–12.8 ka).

14

Speleothems and tufas in arid areas

*Dans les champs de l'observation, le hasard ne favorise que
les esprits préparés.*

In the field of observation, chance favours the prepared mind.

Louis Pasteur (1822–1895)

Lecture, 1854

14.1 Introduction

The term *speleothem* is a hybrid word derived from the two Greek words for ‘cave’ and ‘deposit’, which is slightly misleading, since not all cave deposits are speleothems, but all speleothems are cave deposits. For the sake of simplicity, we define speleothems as secondary mineral deposits formed within caves and usually consisting of calcite or aragonite (calcium carbonate: CaCO_3) but sometimes made up of gypsum (calcium sulfate: CaSO_4). Speleothems thus comprise the well-known dripstone deposits termed *stalactites* (growing down from the roof of the cave) and *stalagmites* (growing up from the floor of the cave, as well as more extensive *flowstone* sheets and curtains (Figure 14.1). Other types of cave deposits include material that has been blown or washed in, as well as cave breccias formed from roof collapses.

Two other common forms of secondary CaCO_3 deposit are *travertine* and *tufa*, which some authors include as speleothems but which are best kept separate. In his monograph *Travertine*, Pentecost (2005, p. 3) defines *travertine* as: ‘a chemically-precipitated continental limestone formed around seepages, springs and along streams and rivers, occasionally in lakes and consisting of calcite or aragonite, of low to moderate intercrystalline porosity and often high . . . framework porosity within a vadose or occasionally shallow phreatic environment’. (The *vadose* zone in caves and bedrock refers to the zone above the regional watertable, while the *phreatic* zone lies beneath the watertable.)

In the third edition of their *Glossary of Geology*, Bates and Jackson (1987, p. 705) define *tufa* as ‘a chemical sedimentary rock composed of calcium carbonate, formed by

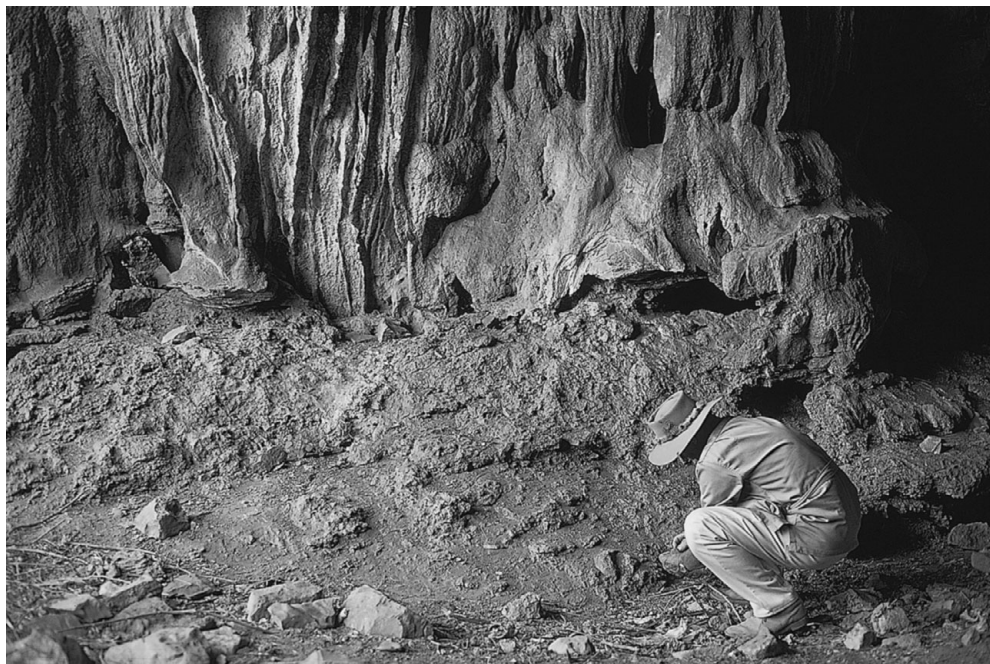


Figure 14.1. Flowstone overlying cave breccia with Middle Stone Age fossils, Porc Epic Cave, Dire Dawa, Ethiopia.

evaporation as a thin, surficial, soft, spongy, cellular or porous, semifriable incrustation around the mouth of a hot or cold calcareous spring or seep, or along a stream carrying calcium carbonate in solution, and exceptionally as a thick, bulbous, concretionary or compact deposit in a lake or along its shore'. They also note that 'it may also be precipitated by algae or bacteria' and comment that 'the hard, dense variety is travertine'. Given that there is no clear genetic distinction between travertine and tufa, we will use the general term tufa to include both, but will consider speleothems as distinct entities.

The actual presence of well-developed cave systems in areas that are now arid indicates that conditions were once wetter when they formed, provided that they had originated above the level of the regional groundwater-table. Caves formed below the watertable and subsequently exposed as a result of tectonic uplift do not necessarily connote wetter conditions at the surface. Because it is not always possible to ascertain how a cave originated, the depositional contents of caves have been the focus of more recent paleoclimatic enquiry and not the caves themselves, although the processes responsible for cave formation remain a major theme of geomorphic research (Ford and Williams, 1989; Ford, 2006; Harmon and Wicks, 2006).

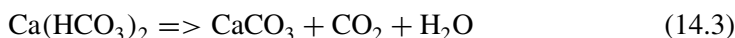
The aim of this chapter is to show how the presence of speleothems and tufas in presently arid and semi-arid areas can provide reliably dated quantitative information

on past changes in precipitation (amount and origin), temperature and vegetation in desert regions during the last half-million years.

14.2 Extracting climatic signals from speleothems

McDermott (2004) and Fairchild et al. (2006) have provided comprehensive reviews of the technical aspects of speleothem analysis, including the precautions that should be observed when attempting to interpret past environmental fluctuations from isotope analysis. During the past few decades, technical improvements in uranium-series dating ($^{230}\text{Th}/^{234}\text{U}$ and $^{234}\text{U}/^{238}\text{U}$) involving the use of inductively coupled plasma mass spectrometers (see Chapter 6) have greatly improved the precision and accuracy of age determinations of stalactites and stalagmites, giving precise ages for when such speleothems were actively forming and when they were not forming in a variety of caves throughout the desert world, dating back to at least 500 ka.

Three fundamental processes are involved in speleothem formation. First, groundwater or water percolating from the surface is enriched in dissolved carbon dioxide to form carbonic acid (Equation 14.1). The carbonic acid attacks the cave limestone, which then forms calcium bicarbonate in solution (Equation 14.2). Finally, the dissolved carbon dioxide comes out of solution within the cave, causing the dissolved calcium carbonate to be precipitated as some form of dripstone, including stalactites and stalagmites (Equation 14.3).



Initial work on speleothems in desert caves where they were no longer actively forming was confined to establishing the times when they were actually formed, because this implied that conditions were more humid then. A major technical breakthrough, pioneered by Hendy (1971) in New Zealand, involved the use of $^{18}\text{O}/^{16}\text{O}$ isotopic ratios measured in speleothem samples to determine past fluctuations in cave air temperature. Hendy also provided a useful set of criteria, still in use today, to test for any significant evaporative effects, because these would affect the temperature interpretation. Later work enlarged the analytical repertoire and included the analysis of $^{12}\text{C}/^{13}\text{C}$ fluctuations as a measure of changes in biological activity and plant cover. Schwarcz et al. (1976) went on to show that analysis of the deuterium/hydrogen (D/H) ratios in fluid inclusions within the dripstone calcite could be used to provide precise measures of past temperature rather than just relative fluctuations from colder to warmer or warmer to colder. These three sets of analyses ($^{18}\text{O}/^{16}\text{O}$, $^{12}\text{C}/^{13}\text{C}$ and D/H ratios) have been at the forefront of speleothem studies ever since. We illustrate this work using a few selected examples drawn from

different parts of the desert world. The aim once again is to be illustrative rather than encyclopaedic.

14.3 Speleothem studies from the desert world

14.3.1 Speleothem studies from peninsular Arabia

We saw in [Section 14.2](#) that increased precipitation in the past was reflected in more active formation of speleothems. Fleitmann et al. (2003a; 2003b; 2007; 2011; and Fleitmann and Matter, 2009) have analysed the oxygen isotopes and fluid inclusions preserved within speleothems from northern Oman and southern Yemen to obtain a record of past wet phases extending back to 330 ka. At Hoti Cave in northern Oman, they found that speleothem deposition was rapid at 6.3–10.5, 78–82, 120–130, 180–200 and 300–330 ka (Fleitmann et al., 2003a; Fleitmann and Matter, 2009). Analysis of the D/H ratios (δD) and the $\delta^{18}O$ values indicated that speleothem deposition coincided with interglacial or interstadial conditions during which groundwater was primarily recharged from moisture derived from a southern source (the Indian Ocean), at a time when the monsoon rainfall belt extended further north and reached northern Oman. Later work on Mukalla Cave in southern Yemen demonstrated that, just as at Hoti Cave in Oman, speleothems grew only during peak interglacial phases such as Marine Isotope Stages (MIS) 1 (early to mid-Holocene), 5a, 5c, 5e, 7a, 7e and 9, with the highest precipitation over that time span coinciding with the last interglacial (MIS 5e) and the lowest occurring during the early to mid-Holocene (Fleitmann et al., 2011). These ages are consistent with ages obtained from four generations of lake deposits in southern Arabia dating to approximately 125 ka, 100 ka, 80 ka and early Holocene (Rosenberg et al., 2011), incidentally also providing circumstantial support for the notion of multiple dispersals of humans from Africa across southern Arabia. Aridity prevailed in southern Arabia between around 75 ka and 10.5 ka, creating a desert barrier for human movement at that time.

Although the speleothem record from peninsular Arabia can document past changes in climate, especially precipitation, it cannot explain the wider causes of such changes. To do this, we need additional lines of evidence. Marine sediment cores from the north-eastern Arabian Sea have shown that laminated bands rich in organic carbon reflect strong, monsoon-induced biological productivity and coincide with relatively warm interstadial events evident in North Atlantic marine sediment cores, as well as in Greenland ice cores, where they are well-known as the Dansgaard-Oeschger warm interstadial events (Schulz et al., 1998). Conversely, during periods when the south-west monsoon was less active, bioturbated bands low in organic carbon accumulated in the Arabian Sea and were synchronous with colder, high-latitude events and associated pulses of meltwater discharge into the North Atlantic, known as Heinrich events. There thus appears to be a correlation between high-latitude temperature changes, variations

in North Atlantic oceanic circulation and low-latitude monsoon activity, although the precise causal links remain a subject of further investigation.

14.3.2 Speleothem studies from the Negev Desert and adjoining region

The Negev Desert of Israel forms a link between the Sahara Desert to the west and the Arabian Desert to the east. The Sahara-Arabian desert belt is one of the driest regions on earth, with less than 50 mm of annual precipitation, although such averages can be misleading given the very high variations in precipitation from year to year. East of the Negev is the Dead Sea Rift, with the hyper-saline Dead Sea in the central portion of the Rift and the Arava Valley lying between the Dead Sea to the north and Aqaba in the eastern tip of the northern Red Sea to the south.

Tufa deposits in the Arava Valley and speleothems in the Negev Desert were forming sporadically during past interglacials (Waldmann et al., 2010). Such depositional events would have been a result of southerly incursions of moist air, given that northern Israel, including the former freshwater Late Pleistocene Lake Lisan in the Dead Sea Rift (Chapter 11), was low during interglacial times but high during glacial times, when there was a greater influx of moist air masses from the eastern Mediterranean during winter (Enzel et al., 2008). One consequence of these short-lived moist phases in the Negev Desert would have been the facilitation of human movement out of Africa into Eurasia via the Levantine Corridor (Vaks et al., 2007; Waldmann et al., 2010), which is discussed in Chapter 17. Lazar and Stein (2011) found evidence of extensive recrystallization of coral reefs along the Red Sea shores as a result of widespread freshwater spring discharge around 140 ka, indicating that humans could have moved along the Red Sea coast at the start of the last interglacial with ready access to good water.

A substantial body of work by Matthews, Vaks, Bar-Matthews and their colleagues has shown considerable geographical variation in speleothem activity across Israel (Matthews et al., 2000; Vaks et al., 2003; McGarry et al., 2004; Vaks et al., 2006; Vaks et al., 2007; Affek et al., 2008; Vaks et al., 2010). Matthews et al. (2000) studied the D/H ratios in fluid inclusions in two fossil speleothems in Soreq Cave south-west of Jerusalem and found that cooler, less evaporative conditions were prevalent over the eastern Mediterranean during glacial times, a finding also confirmed by McGarry et al. (2004). Using ‘clumped isotope’ thermometry (discussed in Chapter 7), Affek et al. (2008) were able to show that temperatures in Soreq Cave were 6–7°C cooler than they are today during the LGM and 3°C cooler at 56 ka.

Analysis of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ variations in speleothems from a number of caves scattered across the southern and central Negev Desert was used to reconstruct past changes in surface vegetation and to determine the probable source of the rainfall associated with speleothem formation. During the past 350 ka, there were major humid periods at 350–310, around 310–290, 220–190 and 142–109 ka, of which all except the

310–290 ka humid phase were interglacial events (Vaks et al., 2010). The wet phases were synchronous with times of sapropel accumulation in the Mediterranean (already discussed in Chapter 10), indicating synchronously wetter conditions over the Sahara and the southern Negev, which facilitated the movement of animals and hominids out of Africa into Eurasia during brief humid episodes (sandwiched between prolonged droughts), the most recent of which was 140–110 ka ago (Vaks et al., 2007). These were synchronous with similar humid phases recorded in speleothems from northern Oman and southern Yemen (Fleitmann et al., 2003a; Fleitmann and Matter, 2009; Fleitmann et al., 2011). However, there was never a simple one-to-one correlation between humid phases in the Negev Desert and interglacials. Vaks et al. (2006) identified major humid intervals in the northern Negev Desert coinciding with glacial phases at 190–150, 76–25 and 23–13 ka and with interglacial phases at 200–190, 137–123 and 84–77 ka. The main source of rainfall in the northern Negev was inferred from $\delta^{18}\text{O}$ values in the speleothems to have been from the eastern Mediterranean, with some possible contribution from tropical southern sources during interglacial episodes. One further interesting insight to emerge from these studies was the influence of rain-shadow effects in eastern Israel, with an effective southward migration of the desert boundary on the eastern flank of the central mountain ridge of Israel during glacial periods and no change relative to the present during interglacials (Vaks et al., 2003).

14.3.3 Speleothem studies from arid and semi-arid Australia

The examples from Israel show that caves that are only a short distance apart can sometimes show quite different responses to precipitation events, indicating that local topographic and hydrologic influences always need to be considered. This is also true of arid southern Australia. For example, a speleothem from Mairs Cave in the arid northern Flinders Ranges of South Australia, dated between 20 and 15 ka, shows peak wetness at 17–16 ka (Cohen et al., 2010b). Quigley et al. (2010b) worked further north in Yudnamutana Gorge and found an increase in humidity after about 11.5 ka, with a peak in wetness at 7–6 ka. One possible interpretation of these seemingly disparate results is that the most northerly of the two sites in the Flinders Ranges reflects the most southerly advance of tropical summer rainfall at that time, while Mairs Cave further south reflects the influence of winter rainfall events. Because no attempt was made to determine the likely precipitation sources, we cannot choose between these two scenarios.

Other speleothem records from Australia have been used in an attempt to elucidate the causes of the late Pleistocene megafaunal extinctions in Australia (Prideaux et al., 2007; Prideaux et al., 2009; Prideaux et al., 2010), and are discussed in detail in Chapter 17. The 185 to 157 ka stable oxygen and carbon isotope record from Victoria Fossil Cave at Naracoorte in semi-arid South Australia revealed that during the 178–162 ka interstadial, regional surface temperatures were anomalously high (possibly

as a result of increased continental area when sea level was lower), and the vegetation at that time was dominated by C_3 plants, in contrast to C_4 grasses during full glacial times. Some recent research involving currently active speleothems in Australia has focussed on comparing seasonal records from instrumentally monitored climatic data with subannual records of trace elements and $\delta^{18}\text{O}$ fluctuations evident in annual speleothem growth rings (Treble et al., 2003; Treble et al., 2005).

14.3.4 Speleothem studies from monsoonal China

The loess record from north-central China extends back more than 2.5 Ma and provides the longest and most detailed continental record of past climatic oscillations for anywhere on earth (Chapter 9). However, although this record is reasonably accurately dated, the error terms are large, and so the loess chronology is far from precise. In order to supplement the more recent portions of the loess record, increasing attention is being paid to the speleothem records from different parts of China. Most of the caves studied so far are not within the drier parts of China, but they are nonetheless very important for our understanding of past climatic changes in the deserts of Asia and elsewhere. There are several reasons for making this assertion. First, the Chinese speleothems are very precisely dated and provide a high-resolution record of past changes in the intensity of the summer and winter Asian monsoon and its two variants, the Indian Monsoon and the East Asia Monsoon, both of which affect the desert margins. Second, the fine resolution allows detailed comparison with both the Greenland ice core records and those from Antarctica, providing a glimpse into possible changes in global atmospheric circulation patterns. Third, the record allows comparison with regions as far distant as north-east Brazil, the Cariaco Basin off Venezuela, peninsular Arabia and the North Atlantic, once again providing a more coherent view of global climatic fluctuations in and beyond the desert margins.

The $\delta^{18}\text{O}$ records from five stalagmites from Hulu Cave near Nanjing spanning the past 75 ka showed that the East Asian monsoon was more intense when Greenland temperatures were warmer and weaker during cold intervals in Greenland (Wang et al., 2001; Wang et al., 2005). The Hulu Cave record was later extended back to the penultimate glacial and deglacial phases, and analysis of three stalagmites demonstrated that both of the glacial terminations occurred in two phases, with an interval of weak monsoon (135.5 to 129.0 ka) followed by an abrupt increase in monsoon strength (Cheng et al., 2006). The overall trend reflected insolation changes related to orbital fluctuations, with some possible influence from changing ice sheet dynamics. Within these broader trends, at least sixteen millennial scale events were evident during the penultimate glacial period, comparable to the Dansgaard-Oeschger cycles of the last glacial period (see Chapter 3). A 224 ka $\delta^{18}\text{O}$ record based on $127\text{ }^{230}\text{Th}$ ages obtained from twelve stalagmites in Sanbao Cave in central China confirmed that the record of changes in the strength of the East Asian monsoon reflects orbitally controlled

variations in high northern latitude insolation, with a 23 ka periodicity punctuated by millennial scale events (Wang et al., 2008). These millennial cycles show a decrease in duration and an increase in frequency during the last two glacial periods, suggesting that they were influenced by changes in ice sheet size. Analysis of $\delta^{18}\text{O}$ from two stalagmites in Dongge Cave located 1,200 km west-south-west of Hulu Cave provide a record of precipitation changes over the past 160 ka consistent with that from Hulu Cave and show that the Last Interglacial Monsoon began quite abruptly at 129.3 ± 0.9 ka and ended equally abruptly at 119.6 ± 0.6 ka, in accord with changes in temperature in the North Atlantic region recorded in Greenland ice cores (Yuan et al., 2004).

A recurrent question in regard to past monsoonal activity in China is whether or not the East Asian summer monsoon (EASM) and the Indian summer monsoon (ISM) were in or out of phase. The EASM is controlled by winds blowing from the south-east flowing across the western Pacific and the South China Sea into eastern China and central China. The ISM is controlled by winds blowing from the south-west across the Indian Ocean and the Bay of Bengal into southern China. A high-resolution $\delta^{18}\text{O}$ record from a stalagmite dated between 53 and 36 ka from Xiaobailong Cave in south-east China shows that millennial scale variations in the ISM revealed in this cave were indeed synchronous with those from Hulu Cave in the path of the EASM (Cai et al., 2006). In addition, some features of the ISM recorded at Xiaobailong Cave showed a negative correlation with the Byrd Ice Core record from Antarctica, consistent with the ‘polar seesaw’ hypothesis of Blunier and Brook (2001).

An intriguing feature to emerge from a $\delta^{18}\text{O}$ record spanning the last 1,810 years comes from Wanxiang Cave in semi-arid north-west China, situated between the Qinghai-Tibetan Plateau to the south-west and the Chinese Loess Plateau to the east, and it concerns the correlation between monsoon variability inferred from the speleothem record and dynastic cultural history as portrayed in written archives (Zhang et al., 2008). The monsoon in this locality was strong during Europe’s Medieval Warm Period but weak during Europe’s Little Ice Age, as well as during the final decades of the Tang, Yuan and Ming dynasties, when popular unrest was widespread as a result of poor harvests. In contrast, the opening decades of the Northern Song Dynasty were times of enhanced summer monsoon activity, increased rice cultivation and population increase. Also evident in this speleothem record was a link between solar variability and the summer monsoon, with weaker monsoons coinciding with times of decreased solar intensity. A comparison between the $\Delta^{14}\text{C}$ record from tree rings, which mainly reflects changes in solar activity, and the $\delta^{18}\text{O}$ record from Hoti Cave in northern Oman also revealed a strong correlation between solar variability and the Indian Ocean monsoon intensity during the wet interval at 9.6–6.1 ka (Neff et al., 2001).

Although careful analysis of Chinese speleothems has provided a more detailed and precise record of past climatic fluctuations in eastern Asia than would otherwise

have been possible (Henderson, 2006), interpretation of those records will always contain elements of uncertainty. For example, Maher and Thompson (2012) have recently argued that the $\delta^{18}\text{O}$ records from Chinese stalagmites do not in fact reveal changes in the amount of summer rainfall but rather reflect changes in the source of the moisture. Because changes in air mass precipitation sources will in any event result in changes in the amount of rainfall, it may be that the differences in interpretations will turn out to be more apparent than real.

14.3.5 Speleothem studies from semi-arid North America

The $\delta^{13}\text{C}$ record from two stalagmites from a cave in central Missouri show positive excursions at 3.5 ka and 1.2–0.9 ka. These episodes are broadly coincident with dry intervals inferred from other lines of evidence across the semi-arid Great Plains of North America (Denniston et al., 2007). Possible factors controlling the $\delta^{13}\text{C}$ excursions include a change in the abundance of C_4 plants above the cave and/or a greater input from bedrock carbon, both of which could lead to a reduction in effective moisture. Given that the $\delta^{18}\text{O}$ record showed no change during the times of anomalous $\delta^{13}\text{C}$, it would appear that there was no change in mean annual temperature or in rainfall seasonality during these episodes of inferred aridity.

A particular form of speleothem known as ‘cave mammillaries’ has been used to determine past elevations of the watertable at nine sites in the Grand Canyon (Polyak et al., 2008). Using uranium-lead dating of the speleothems and equating rates of groundwater-table decline with concomitant rates of incision, these authors found that in the western Grand Canyon, the rates amounted to 55–123 m/Ma during the past 17 Ma. In the eastern Grand Canyon, the corresponding rates were much faster, amounting to 166–411 m/Ma. The overall conclusion was that the Grand Canyon has evolved through headward erosion from west to east, accompanied by accelerated incision in the eastern sector during the past 3.7 Ma or so.

14.4 Extracting climatic signals from tufas

Tufas form in much the same way as speleothems. Carbon dioxide comes out of solution in river, lakes or spring water, often as a result of an increase in water temperature, and the dissolved calcium carbonate is precipitated (Equation 14.3). This tends to occur at the outlet of springs, on waterfalls and along lake, swamp or even flood-plain margins. Tufas occur in a wide variety of locations within desert landscapes. They may form a series of roughly horizontal benches along valley sides, especially those associated with former river terraces and valley fills (Butzer and Hansen, 1968; Butzer, 1984; Pentecost, 2005; Pedley, 2009). Tufa dams are common in ephemeral stream channels, particularly near waterfalls and knick points, as in central Afghanistan (Bouyx and Pias, 1971) and the seasonally wet Kimberley region

of north-west Australia (Wright, 2000). If we accept the radiocarbon ages obtained directly from tufa samples, it would appear that the tufa dams in central Afghanistan were formed between >40 ka and 25 ka and again between 14.2 ka and 11.6 ka, or before and after the LGM. Dissolution of the calcareous bedrock in the headwaters may have been inhibited by glacial aridity (Bouyx and Pias, 1971).

Many former springs are demarcated by tufa mounds, of which the mound springs in the Great Artesian Basin of central Australia are perhaps the most striking, with individual tufa deposits forming flat-topped hills tens of metres in height with OSL ages ranging back into the Middle Pleistocene (Prescott and Habermehl, 2008). Tufa beds are often intercalated within fluvial and lacustrine sediments, and may contain the remains of freshwater gastropod shells (Abell and Williams, 1989) and ostracod valves (De Deckker and Williams, 1993), or even fossil reed stems (Williams et al., 2001). Many tufas precipitate out as a result of biological activity involving cyanobacteria, heterotrophic bacteria and diatoms (Pedley, 2009). Tufa benches formed from algal limestone crop out around a terminal Pleistocene high shoreline of the shallower of the two lakes within the Deriba Caldera of Jebel Marra volcano in arid north-west Sudan (Williams et al., 1980). Many prehistoric sites in the drier parts of southern and eastern Africa are associated with tufa deposits located at the mouth of previously active springs (Clark and Williams, 1977; Williams et al., 1977; Butzer, 1984). Some tufa deposits form underwater in lakes, either as a result of biological activity or in association with hot springs. In Lake Abhe in the Afar Desert, a striking series of biogenic limestone pillars (Figure 14.2) bear witness to a time when this now very shallow saline lake was deep and fresh (Fontes and Pouchan, 1975). In hyper-saline lakes, such as the Coorong Lakes of South Australia, cyanobacterial mats cemented by carbonate form stromatolites (Mee et al., 2007) similar in form to those that formed in Precambrian lagoons more than a billion years ago. Tufas can also form when freshwater lakes dry out in arid or semi-arid areas and are common in many Cenozoic formations in North America and around the Mediterranean (Alonso-Zarza et al. (2006).

Although they are visually striking features of certain desert landscapes, tufas/travertines are not especially useful for reconstructing past changes in climate with any degree of precision. There are several reasons for this. Owing to their highly porous fabric, uranium is easily leached from the parent tufa, so that uranium-series dating is problematic. In addition, tufas are exposed to accretions of detrital thorium brought in from wind-blown dust, and this further complicates dating. Furthermore, the location of many tufas is controlled by topographic and tectonic factors, such as the presence of springs along faults or at breaks of slope, so unravelling any climatic influence is unrewarding. Spring tufas associated with changes in the elevation of the groundwater-table will only provide a blurred climatic signal, and dating such deposits runs into the problem of contamination from dead carbon. In certain situations, this problem can be overcome. For example, analysis of the $\delta^{18}\text{O}$ fluctuations revealed



Figure 14.2. Algal limestone pillars formed when Lake Abhe was full during the late Pleistocene and early Holocene, Afar Desert, Ethiopia. (Photo: Françoise Gasse.)

in annually banded modern and mid-Holocene travertines/tufas in the Grand Canyon of Arizona suggests that they were formed as a result of monsoon-generated summer floods during the mid-Holocene, just as they are today (O'Brien et al., 2006).

Brook et al. (1997) compared the radiocarbon and U-series ages obtained from speleothems and tufas from the summer rainfall zone of southern Africa (Namibia, Botswana, northern Cape and the Transvaal) with those obtained on similar deposits across Somalia. They found that during the last 250 ka, when conditions were wet in southern Africa they were dry in Somalia, and conversely. In Somalia, speleothem, tufa and rock shelter sediments indicated wetter conditions in this arid region at 260–250, 176–160, 116–113, 87–75, 13, 10, 7.5 and 1.5 ka. In southern Africa, conditions were

wet during late glacial times and dry during the early Holocene. Southern Africa was apparently wetter at 202–186, 50–43, 38–35, 31–29, 26–21 and 19–14 ka. Earlier work in this region showed substantial uncertainties associated with some of the younger ages based on radiocarbon dating, so some recalibration was deemed necessary.

In north-east Brazil, Wang et al. (2004) used a combination of speleothems and tufa deposits to provide a record of wet phases during the past 210 ka. They found that wet periods in this presently semi-arid region coincided with times of weak East Asian summer monsoons in China, cold periods in Greenland, phases of iceberg discharge in the North Atlantic (Heinrich Events) and reduced run-off into the Cariaco Basin off the coast of Venezuela. They concluded that the wet intervals probably reflected southward displacement of the Intertropical Convergence Zone, leading to forest expansion and the creation of a forest corridor between the Atlantic and Amazonian rainforests. Earlier work on speleothems and travertines/tufas that date back to around 400 ka in semi-arid north-east Brazil (Auler and Smart, 2001) had shown that during the LGM, the regional water-table was 13 ± 1 m higher than it is today (and even higher during MIS 6), in contrast to pollen evidence of a drier glacial climate in Amazonia and southern Brazil. These studies demonstrated that there was considerable regional variation in precipitation in this region during glacial times.

14.5 Conclusion

Technical advances in extracting climatic signals from speleothems pioneered in the 1970s were strengthened thanks to improvements in dating methods perfected during the late 1980s and onwards. These allowed very precise ages to be obtained from cave stalagmites and have helped extend the practical dating range back to about half a million years. Analysis of the stable isotopes of oxygen and carbon has allowed past changes in precipitation and in surface plant cover to be determined with reasonable quantitative accuracy. Additional analysis of fluid inclusions within speleothems, notably of the ratio of deuterium to hydrogen, has allowed an independent estimate of past temperature. The high-resolution records from Israel, peninsular Arabia, north-east Brazil and China have provided invaluable insights into possible links between summer monsoon fluctuations, temperature changes in the North Atlantic and, at finer time scales, changes in solar intensity. However, it is worth sounding a note of caution. An annually laminated stalagmite from Mechara just south of the Ethiopian Rift Valley in south-east Ethiopia grew for around 440 years from 5,023 yr BP onwards (Asrat et al., 2007). Comparison of different climate proxies, such as $\delta^{18}\text{O}$, $\delta^{13}\text{C}$, $^{234}\text{U}/^{238}\text{U}$, annual growth rate and the fluorescence index, revealed a variety of responses to a single climate forcing, confirming the need for caution in interpreting climate signals from speleothem records.

Comparisons between precise speleothem chronologies of certain critical episodes, such as the Younger Dryas cold event and the Bølling/Allerød warm interstadial phase,

that are evident in the speleothem data and those recorded in Greenland has shown subtle differences in timing and intensity in areas such as Tunisia and southern France (Genty et al., 2006). Another example will suffice to show the versatility and wide applicability of speleothem research. The very precise chronology now available from the Antro del Corchia Cave speleothems in north-west Italy has allowed millennial scale cold events evident in North Atlantic marine cores to be precisely dated for the first time, namely, to 112–109 ka and 105–102.6 ka (Drysdale et al., 2007). Speleothem research is of far-reaching global significance, and an invaluable adjunct to less exactly dated studies of desert dunes, lakes, rivers, dust and fossils. It is therefore no surprise that in his review of karst geomorphology, caves and cave deposits, Ford (2006, p. 9) commented on ‘a worldwide speleothem paleoclimate bandwagon in environmental sciences today’.

15

Desert soils, paleosols and duricrusts

Excellent soil scientists for some kinds of research, including detailed soil surveys, fail utterly in reconnaissance soil mapping. They may be unable to visualize large and complex patterns and become mentally harassed by indecision in the face of vague and apparently conflicting evidence.

U.S. Department of Agriculture
Soil Survey Manual, Agricultural Handbook 18
(1951, pp. 437–438)

15.1 Introduction

The three most important natural resources in all drylands (and, indeed, in less arid areas) are water, soils and vegetation. Soils in the arid world differ from their humid counterparts in being generally low in organic matter. In addition, they are often saline and/or alkaline (Williams, 1968b; Amit and Yaalon, 1996). Both of these factors limit plant growth and constrain human use of desert soils. The broad distribution of such soils is reasonably well-known both globally (FAO, 1991; IUSS Working Group WRB, 2007), on a continental scale (e.g., McKenzie et al., 2004; Soil Survey Staff, 2010) and on a national scale (e.g., Chinese Soil Taxonomic Classification, 1991), although, of course, all soils mapping is provisional, as new methods of remote sensing and geochemical mapping improve apace.

Within the drylands, not all soils are currently active and in balance with the present-day bioclimatic environment. Scattered across many deserts, there are sporadic remains of former soils, as well as more resistant formations cemented with carbonate, iron, silica or gypsum that are termed *duricrusts*. The relict soils are variously referred to as fossil soils, or *paleosols*, implying that they are no longer forming (Yaalon, 1971; Birkeland, 1999; Retallack, 2001). The presence of such soils in hyper-arid areas indicates that the climate was once wetter than it is today (Khadkikar et al., 2000), because soils need both moisture and biological activity in order to develop.

Using soils to infer past climate is a difficult task, given that soil formation involves a number of independent influences or soil forming factors, only one of which is climate. The aim of this chapter is to consider the utility and disadvantages of using soils and duricrusts to identify past climatic changes in deserts and their margins. Before doing so, it is necessary to discuss the factors of soil formation and briefly to review some of the diagnostic characteristics of so-called desert soils and soils found in what are now deserts. We begin by defining what a soil is.

15.2 What is a soil?

Given the understandable but sometimes misleading emphasis among soil scientists on describing the two-dimensional soil profiles revealed in soil pits and the one-dimensional view afforded by soil augers, it is easy to forget that soils are in fact three-dimensional features of the landscape (Birkeland, 1999; Retallack, 2001). More than half a century ago, the *Soil Survey Manual* (1951, p. 7) pointed out that soils are ‘landscapes as well as profiles’ and emphasised that ‘soil is the natural medium for the growth of land plants, whether or not it has “developed” soil horizons’. The *Manual* went on to define soil as ‘the collection of natural bodies occupying portions of the earth’s surface that support plants and that have properties due to the integrated effects of climate and living matter, acting upon parent material, as conditioned by relief, over periods of time’ (op. cit., p. 8). This definition remains just as valid today, and it will be the one adopted here.

In this comprehensive definition, the five classic factors of soil formation are clearly identified: parent material, topography, biological activity, climate and time (Jenny, 1941; Paton, 1978, pp. 96–108). In the early stages of soil development, parent material and topography exert the greatest influences on soil characteristics, and they may be considered as relatively passive factors of soil formation (Paton, 1978, p. 96) (Figure 15.1). As a result, we can recognise the varying influence exerted by different igneous, metamorphic and sedimentary rocks on the physical and chemical characteristics of immature soils that developed, for example, on the more common rocks, such as granite, basalt, sandstone, siltstone and limestone. In some cases, soils may occur at the top of a deep weathering profile, while in other instances, there may be a series of buried soils within successive layers of alluvium, loess and volcanic ash or lava flows. Many soils found in deserts are formed on relatively young sediments, including alluvial and lacustrine parent materials, as well as eolian deposits such as sand and desert dust or loess. Sequences of buried soils are common in such environments.

Topography plays an important role in soil development in several ways (Milne, 1936). Water percolates through weathered rock laterally as well as vertically, and is a potent agent of mechanical eluviation of finer soil particles and their redeposition along the base of hills and inselbergs to form an aureole of finer-textured soils, as in

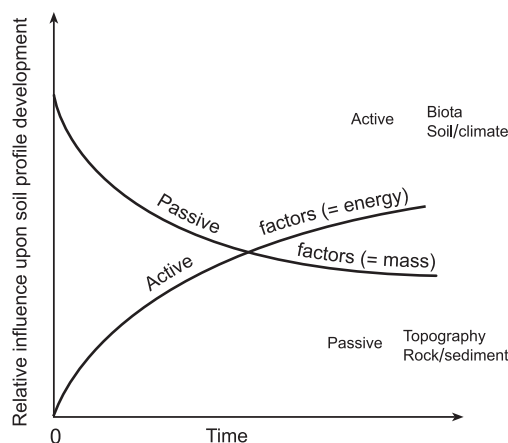


Figure 15.1. Factors of soil formation.

semi-arid central Sudan (Ruxton, 1958). Milne (1935; 1936; 1947), Ruxton (1958) and Webster (1965) identified and mapped soils forming genetically linked topographic patterns, or *toposequences*, in semi-arid East Africa, Sudan and Zambia. Such toposequences are also known as soil *catenas*, from the Latin word for chain (Milne, 1935). It is worth remembering that a number of different soil types can occur within one catena, ranging from weakly developed regosols to highly organic histosols (see Table 15.1). Subsequent erosion may only leave remnants of one soil type, so great caution is needed before invoking a particular regional climate, especially if much of the former topography has been destroyed.

As time progresses, the influence of the more active factors (climate; biological activity) will tend to outweigh that of the passive factors (Figure 15.1). In deserts, where water is a limiting factor, biological activity is curtailed and only becomes important during those intervals of time when precipitation increases, that is, during wetter climatic phases. If the humid climatic phases are brief, little soil development will occur other than some minor organic staining and weak soil horizon development of the parent material, whether it be weathered bedrock or sediment. Strictly speaking, climate should refer to *soil climate*, and will reflect the interaction of soil water and temperature, modulated by soil permeability and infiltration capacity.

The influences of soil climate and local topography are mediated through the biota, including soil microorganisms, insects (especially termites, ants and worms), larger burrowing mammals and birds. The interaction of each of the agents of soil formation produces what are known as soil horizons, in which a horizon is a gently sloping or horizontal layer of soil with certain diagnostic traits, such as colour, structure (the arrangement of soil aggregates, or peds), texture (clay, silt and sand content as assessed in the field by manipulating moistened soil) and other features, such as calcium

Table 15.1. *Major international soil groups recognised in drylands. (Modified from IUSS Working Group WRB, 2007.)*

Soil Group	Characteristics
Andosols	Soils formed in volcanic ash, with abundant volcanic glass (Andisols)
Arenosols	Sandy soils with minimal texture contrast and weak or no soil horizons (Entisols)
Cambisols	Soils with weakly developed horizons, depleted of bases and/or iron and aluminium but with some weatherable minerals; occur mainly in humid and subhumid areas (Inceptisols)
Chernozems	Soils formed under grassland in semi-arid to subhumid areas; rich in humus, bases and calcium carbonate (Mollisols)
Ferralsols	Soils rich in iron or aluminium (Ultisols)
Fluvisols	Young soils in alluvial deposits still showing signs of alluvial stratification (Entisols)
Gleysols	Soils that have been permanently or sporadically waterlogged; grey or green colours; ironstone nodules may occur in the subsoil
Gypsisols	Soils of arid areas with secondary accumulations of gypsum (calcium sulfate)
Histosols	Highly organic soil with at least 20–30 per cent organic matter by weight in a layer at least 40 cm thick (Histosols)
Kastanozems	Red-brown, humus-rich soils of semi-arid steppes and grasslands (Mollisols)
Lithosols	Shallow soils, mainly composed of unweathered rock fragments
Luvisols	Texture-contrast soils with higher clay content in the subsoil; the eluvial (Ae) horizon overlies an illuvial (Bt) horizon enriched in clay
Planosols	Soils with a bleached, light-coloured, near-surface horizon over a horizon with a higher clay content (Ultisols)
Podzols	Ashy grey acidic soils formed on sands; strongly leached surface layer; subsurface accumulation of humus mixed with amorphous iron and/or aluminium (Spodosols)
Rankers	Soils developed over non-calcareous rock, with A/C profiles (Entisols)
Regosols	Soils formed on deep, unconsolidated, recently deposited sands or alluvium (Entisols)
Rendzinas	Red clay-rich soils formed on limestone or eolianite, with high calcium carbonate content
Solonchaks	Saline and often alkaline soils of hot arid areas; often formed on playas or sebkhas with salt forming near the surface as a result of strong evaporation
Solonetz	Soils with tough, impermeable hardpan and domed prismatic structure; often saline and alkaline.
Vertisols	Heavy, dark churning clay soils with deep vertical cracking in the dry season; contain abundant swelling clay minerals (notably, montmorillonite); Variable salinity and alkalinity (Vertisols)

carbonate nodules, gypsum crystals or salt content. The use of the letters A, B and C for soil horizons is now universal but was originally applied in Russia specifically to chernozems (Table 15.1) (*Soil Survey Manual*, 1951, p. 176). Soil horizons are defined in Section 15.4.

Table 15.2. *Soil orders of North America. (Modified from Soil Survey Staff, 2010.)*

Soil Order	Characteristics
Alfisols	Arable soils with >3 consecutive months with enough soil water for plant growth
Andisols	Soils formed in volcanic ash, with abundant volcanic glass (Andosols)
Aridisols	Dry, desert-like soils, often rich in calcium carbonate, with low organic content and sparse vegetation cover
Entisols	Soil with little or no profile development and lacking diagnostic soil horizons except for weak A-horizon (Arenosols; Fluvisols; Regosols)
Gelisols	Cold-climate soil with permafrost within 2 m of the surface
Histosols	Highly organic soil with at least 20–30 per cent organic matter by weight in a layer at least 40 cm thick (Histosols)
Inceptisols	Soil with weakly developed horizons, depleted of bases and/or iron and aluminium but with some weatherable minerals; occurs mainly in humid and subhumid areas
Mollisols	Soils formed under grassland in semi-arid to subhumid areas; rich in humus, bases and calcium carbonate (Chernozems; Kastanozems)
Oxisols	Thick weathered soils of the humid tropics; mostly depleted of unweathered minerals; red to yellow colours (Ferralsols)
Spodosols	Ashy grey acidic soils formed on sands; strongly leached surface layer; subsurface accumulation of humus mixed with amorphous iron and/or aluminium (Podzols)
Ultisols	Weathered red/yellow clay-rich acidic soils low in bases (Ferralsols)
Vertisols	Heavy, dark churning clay soils with deep vertical cracking in the dry season; contain abundant swelling clay minerals (notably, montmorillonite); variable salinity and alkalinity (Vertisols)

15.3 Soil classification

During the late nineteenth and early twentieth centuries, the great pioneering soil scientists Dokuchaev, Marbut and their co-workers mapped the soils of the drier regions of Russia and North America at a broad reconnaissance level. In Russia, the soils were aligned roughly parallel to latitude, giving rise to the notion that climate was the primary control over soil distribution, with podzols in the colder north and chernozems in the warmer, somewhat wetter south. North America is aligned in a different way, with the Rockies separating west from east and producing a pronounced rain-shadow effect in the Great Basin. Early soil mapping defined a zone of pedalfers in the more humid east and pedocals in the drier west, with pedalfers being soils rich in aluminium and/or iron and pedocals being soils rich in calcium carbonate. Since that time, soil classification has greatly progressed, and in North America, twelve Soil Orders have now been recognised (Soil Survey Staff, 2010 and Table 15.2).

Efforts to secure an international classification began with the FAO-UNESCO Soil Map of the World at 1 in 25 million scale and its successive versions published between 1971 and 1981. From 1982 to 1991, this task was delegated to the International

Reference Base for Soil Classification, with input from UNEP and the International Union of Soil Sciences, culminating in the current World Reference Base for Soil Resources (IUSS Working Group WRB (2007)). It is interesting to compare Table 15.1 distilled from the work of this group to the soils recognised in North America (Table 15.2) by the Soil Survey Staff (2010) of the U.S. Department of Agriculture.

For ease of comparison, the equivalent or near-equivalent soils are listed in brackets in Tables 15.1 and 15.2. Certain of the Soil Orders recognised in North America do not feature as such in the international classification (e.g., alfisols, gelisols, inceptisols), just as certain of the international Major Soil Groups are not identified at this broad level in the North American classification system, although they undoubtedly do occur in North America (e.g., solonetz, solonchaks, rendzinas, kastanozems). Table 15.1 is not the full list of thirty-two Major Soil Groups, but specifies the twenty soils most commonly found in deserts and their margins, whether they are active or fossil. The list of characteristics is gleaned from the key references given in the previous sections, as well as the author's own experience of mapping and describing these soils in the drier parts of Africa, Australia and Asia.

There has long been dispute among soil scientists as to whether soils should be classified purely on the basis of observable morphology and physical field attributes or whether they should be classified using both physical and chemical criteria. The Australian soil scientist Keith Northcote developed what he described as a Factual Key for classifying soils in the field (Northcote, 1971). He identified three broad categories of soil profile on the basis of field soil texture: uniform, gradational and texture-contrast. These then formed the basis for more detailed subdivision. In North America and Europe, preference was given to including chemical properties in any classification system. For example, the once widely popular Seventh Approximation at a soil classification (Soil Survey Staff, 1960) used a combination of chemical and physical criteria to define soil types. The Seventh Approximation suffered from using unduly abstruse soil names concocted from a hybrid of Greek and Latin roots to produce such forbiddingly opaque names as *Natrargidic Mazustert*, which simply means an alkaline clay with a surface crust, and *Orthic Grumustert* for a non-saline, non-alkaline cracking clay with a self-mulching surface. Aside from the obscure and formidable terminology, which is really only a minor irritant, a fundamental weakness of the Seventh Approximation, and of certain of its successors, is the uncritical use of climatic zonalism as a genetic basis for classifying soils. The next section explains this key point in more detail.

15.4 Soil-forming processes

Well-developed soil horizons have long been considered as diagnostic of mature soil profiles (Joffe, 1949; Kubiëna, 1950; Mohr and van Baren, 1959; Mohr et al., 1972; Buol et al., 1973; Duchaufour, 1978; Paton et al., 1995; Birkeland, 1999;

Retallack, 2001). Following the early broad-scale mapping of soils in Russia, North America and Europe (summarised by Joffe, 1949, and by Kubiëna, 1950), many soil scientists concluded that climate was *the* dominant factor in soil formation, and so a zonal system of soil classification developed in which certain soils were considered to be diagnostic of certain climates. For instance, lateritic soils were widely believed to require a hot, wet tropical climate for their formation, and podzols were believed to require a cool, wet climate characteristic of high northern latitudes. Difficulties arose once podzols were found in tropical regions in association with sandy parent materials and laterites were found actively forming at the present time on ultra-basic rocks in temperate latitudes (Paton and Williams, 1972). In the case of lateritic soils, parent material plays a major role, as does an efficient leaching regime, as recognised by Milne three-quarters of a century ago (Milne, 1938, 1947).

A second problem relates to the relative importance of vertical as opposed to lateral processes in the development of soil horizons. The classic subdivision of soil profiles was (and is) into A, B and C or R horizons, with some lower case suffix to qualify the horizon further, such as a B_{ca}-horizon rich in carbonate. The A-horizon was the generally more organic surface horizon or top-soil, often with less clay content than the underlying B-horizon. The A- and B-horizons form the solum. The C-horizon was the slightly modified parent material in which the soil had formed, and R referred to unaltered bedrock. Traditionally, texture contrast soils were thought to have formed by eluviation of finer particles from the top-soil (or A-horizon) and their redeposition as illuvial clay within the subsoil (or textural B-horizon). Sandy top-soils overlying clay-rich B-horizons were thought to reflect the action of vertically operating processes of surface eluviation (clay removal from the A-horizon) and subsurface illuviation (clay deposition within the B-horizon). Brewer (1955) and Oertel (1968), among others, challenged this model and rejected overly facile inferences about illuvial processes in soils. In addition, a growing number of soil scientists became aware of the important role played by insects such as ants, termites and other burrowing organisms in bringing soil material to the surface, whence it is washed down the slope to form a sandy mantle overlying an often clay-rich substrate (Nye, 1954; Nye, 1955; Watson, 1961; Watson, 1962; Watson, 1964; Williams, 1968c; Paton, 1978; Williams, 1978; Johnson, 1993; Paton et al., 1995).

A third and more subtle problem concerns polygenic soil profiles where the soil horizons relate to several vertically stacked soils in which soil-forming processes have continued to operate in both soil profiles. The result is that some of the original properties of the underlying soil are camouflaged or even obliterated by the processes operating in the overlying younger soil, making it hard to interpret the conditions under which the underlying older soil had developed. One potentially useful way of elucidating soil history is to study the microscopic evidence of past soil processes as revealed by subtle changes in the arrangement of soil materials, otherwise known as soil micromorphology.

Brewer (1964) was among the first to establish soil micromorphology as a rigorous scientific discipline, but the specialised names now widely used in this field of soil science are not for the faint-hearted. The techniques of soil fabric analysis have been widely used in soil science, as well as in other disciplines such as geo-archaeology and glacial geomorphology, and they are an invaluable adjunct to field-based soil description, in that the thin sections can show the actual amounts of illuvial clay within a given soil horizon.

15.5 Examples of desert soils

Given the iconic status of desert dunes, it seems fitting to begin with soils developed on and within desert dunes and sand plains, especially since many dune soils are hard to recognise as actual soils. Since 1977, Neil Munro and his colleagues have conducted extensive soil surveys throughout Saudi Arabia, Yemen, Oman, Jordan, Bahrain and the United Arab Emirates, combined with a program of radiocarbon and luminescence dating and archaeological research (Munro et al., 2012). It is worth quoting one of their conclusions:

At the present time on the Yemen Tihama dunes, moderate summer rainfall allows substantial rainfed millet production on mobile dunes (and concomitant dune stabilisation) and thus the requisite for a developed soil horizon to allow a vegetation cover to form is false . . . and one does not have to search for true paleosols with Cambic [i.e., clay-rich] horizons to show that lands were grassed or wooded and supporting [prehistoric human] populations. (op. cit., p. 32)

This comment would equally apply to the sand deserts of Africa, Asia, Australia and the Americas. Such soils would qualify as arenosols (see Table 15.1), which are sandy soils with minimal texture contrast and weak or no soil horizons.

However, many dune soils do show some degree of horizon development, and they often have modest amounts of clay at shallow depths, together with soft calcium carbonate (CaCO_3) aggregates or hard, irregular carbonate nodules and rhizcretions (Williams, 1968a; Williams et al., 1991a; Amit et al., 2007). Rhizcretions are roughly cylindrical or downward-tapering carbonate pipes that formed around tree roots, a process that can be observed on living tree roots today in regions as far apart as Algeria, Sudan, the Thar Desert and Australia. They are especially common in coastal eolianites (Sprigg, 1959; Sprigg, 1979). Dust storms are the primary sources of the clay and calcium carbonate found within dune soils, although some carbonate may have travelled in solution in groundwater. Analysis of the strontium and neodymium isotopes (Chapter 7) within desert soil carbonates is consistent with deflation from distant sources, including continental shelves exposed during times of low glacial sea level (Dart et al., 2007). Carbonate rich soils in varying stages of development are common on pediment surfaces in areas as far-removed as the Mojave Desert in the south-west United States, the southern Negev Desert, the Thar Desert of Rajasthan in

north-west India, the Alashan Plateau of Inner Mongolia in northern China and the Namib Desert.

Along the strategically important railway between Baotou and Hohot in Inner Mongolia, dune encroachment onto the tracks is a perennial problem. To overcome this problem, the Chinese have adopted an ingenious albeit labour-intensive method of dune stabilisation. Rows of dried wheat stalks (or, in the northern Taklamakan Desert, dried reed stems from the piedmont marshes south of the Tian Shan) are pushed well into the sand to form small square hedges about a metre wide and up to 0.5 m high. These ‘chequer-boards’ act as temporary windbreaks, allowing wind-blown dust to accumulate within each square. The dust is washed down into the sand during the sparse summer rains and, within a few decades, a soil enriched in silt and clay forms, soil moisture becomes trapped and seeds of ephemeral and perennial desert shrubs and grasses germinate and stabilise the previously mobile dunes.

Many of the dunes along the southern margins of the Sahara are now vegetated and stable (Grove and Warren, 1968; Talbot, 1980). The dunes have weakly developed A-horizons enriched in silt and clay from many centuries of dust storms. Such soils are very vulnerable to gully erosion during intense rainstorms. One such extreme event occurred during the wet season in mid-1974 near the village of Janjari in central Niger, about midway between Niamey and Agades (Talbot and Williams, 1978; Talbot and Williams, 1979). The fixed dunes in a small area 10–15 km in radius between Tahoua and Abalak were eroded during this storm, and a series of sandy alluvial fans formed along their foot-slopes. Older fan sediments were visible in the banks of a channel incised upstream of one of the most recent fans, and buried soils exposed in the banks showed that previous fans had become vegetated and stable. The youngest buried soil was stratigraphically earlier than charcoal from a fireplace dated to 335 ± 60 yr BP (N-2129). This same soil was found to be widespread throughout the region and may have formed during a slightly wetter interval evident elsewhere in the Sahel and dated to about 150–350 years ago (Talbot and Williams, 1978; Talbot and Williams, 1979). One conclusion to emerge from this work is that episodic dune dissection, fan deposition, soil formation and fan stabilisation have been typical of the last few thousand years, indicating that humans were not responsible for the dune erosion and that desertification processes are reversible.

15.6 Attributes of soils and soil landscapes in semi-arid regions

Desert margins are often characterised by very gently sloping alluvial fans which can occupy vast areas, such as the Gezira alluvial fan between the lower Blue and White Nile rivers (40,000 km²) and the Riverine Plain in the lower Murray-Darling Basin of south-east Australia (77,000 km²). These vast, low-angle, inland alluvial fans have been termed ‘megafans’ by Leier et al. (2005), and they appear to be confined to

climatically restricted areas characterised by a long dry season and a distinct wet season, flanked upstream by often high upland catchments. The landform assemblage of these fans is complex, and includes extensive level clay plains, several generations of former stream channels, sand dunes, closed depressions and back swamps (Chapter 10, Figures 10.3–10.5). Although they display a high degree of geomorphic variability, these regions possess many common attributes. They are areas of low, erratic rainfall and high rates of evaporation. Inputs of cyclic salt have contributed to the naturally high levels of soil and subsoil salinity. When combined with inadequate leaching and drainage, they offer peculiar problems for agricultural development that are exacerbated by the general lack of surface water and the presence of groundwater of variable quality.

In addition, they have undergone a complex depositional history that exerts a powerful control over the distribution of soils and their physical and chemical characteristics. As a result of the variable depositional history of their parent materials, the soils are equally variable in texture, salinity and permeability. Indeed, many of the soils in semi-arid areas have physical and chemical characteristics that have little to do with present-day climatic conditions in those regions (Williams, 1968a; Williams, 1968b). On the contrary, they owe much to their past history, especially the depositional history of their parent materials.

The cracking clay soils of central Sudan illustrate this point very well (Buursink, 1971; Williams et al., 1982; Blokhuis, 1993). Attributed variously to deposition from wind-blown dust or to in situ weathering of the underlying Proterozoic Basement Complex rocks, it was not until the pioneering work of Tothill (1946; 1948) that their alluvial origin was recognised. Tothill based his conclusions on the heavy mineral content of the clay soils, which denoted a volcanic source from the Ethiopian Highlands via the Blue Nile, and on the presence of freshwater gastropod shells found down to depths of 2 m, including the aptly named *Cleopatra bulimoides*. Williams (1966) built on this work by providing the first radiocarbon ages of shells at 1.5–2.0 m depth at two sites adjacent to the present White Nile. Considerable effort has been devoted since then to developing a detailed late Pleistocene and Holocene flood history for both the White and the Blue Nile rivers (Williams and Adamson, 1973; Williams and Adamson, 1974; Williams and Adamson, 1980; Williams et al., 2006c; Williams, 2009b; Williams et al., 2010b). Indeed, some puzzling features of the soil landscapes in the lower White Nile Valley are not explicable without a detailed understanding of its geologically recent history (Williams et al., 2000). These features include a braided channel pattern despite a remarkably low gradient, a flood gradient of 1 in 100,000, the strongly localised presence of highly saline subsoils, the presence of subsurface carbonates and evaporites, the presence of widespread cracking clays overlying quartz sand dunes, the presence of lake and swamp fossil faunas in now arid areas west of the river and the presence of buried shell-beds located up to 4 metres above present-day

mean maximum flood level and up to 10 km from the river. Each of these puzzles has now been explained (Williams et al., 2000; Williams et al., 2006b; Williams et al., 2010b; and Chapter 10).

It is important to realise that non-climatic factors are often involved in soil formation in semi-arid and seasonally wet tropical areas on the margins of deserts. For example, in the seasonally wet tropics of Africa, Australia, South America and Asia, three-layered soils are widespread but defy classification into the conventional A-, B- and C-horizons. They consist of a sandy top-soil (M-horizon), an underlying stone layer (S-horizon) and a highly weathered clay-rich subsoil (W-horizon). Buried stone layers are characteristic of strongly weathered granites containing resistant quartz veins, and are widespread in seasonally wet tropical Asia, Africa, South America and in the Piedmont region of the United States (Parizek and Woodruff, 1956; Vogt and Vincent, 1966; Troitsky et al., 1968; Ségalen, 1969). They have been attributed to soil creep (Ireland et al., 1939; Eargle, 1940), burial of stony colluvium, alluvium or erosional lag gravels (Ruhe, 1956; Parizek and Woodruff, 1957; Ruhe, 1959; Marchesseau, 1967; Fairbridge and Finkl, 1984), slope retreat (Ségalen, 1969), the swelling of clay soils (Jessup, 1960a; Mabbutt, 1965a) and termite activity (Charter, 1950; Nye, 1955; Watson, 1962; Williams, 1968c; Lee and Wood, 1971; Williams, 1978).

Although a variety of mechanisms are capable of producing buried stone layers, one of the most effective is through the activity of mound-building termites (Figure 15.2). Termites (of which there are more than 2,600 species) are related to cockroaches and, thanks to special microorganisms in their gut, can feed almost exclusively on cellulose. In northern Australia, where termites play a major role in savanna ecosystem dynamics, there are more than 150 species of termite (Andersen et al., 2005). Of the four broad groups of termites (wood-feeders, soil-feeders, debris-feeders and grass-harvesters), only the latter need concern us here. The grass-harvesting species collect dry grass, cut it into small pieces and store the pieces in the galleries inside their mounds. Two of the more common mound-building species are *Tumulitermes hastilis* and *Nasutitermes triodiae* (Figure 15.3). *T. hastilis* builds quite small mounds up to 75 cm high that are usually abandoned after about three years, but the mounds of *N. triodiae* can exceed 6 m in height and attain ages of up to 100 years, although less than half that age is more common. Williams (1968c; 1978) monitored rates of soil erosion by slopewash and soil creep on granite hill slopes in seasonally wet tropical northern Australia and found that current rates of surface lowering are balanced by top-soil replenishment from termite mound breakdown and redistribution across the slope. At existing termite mound densities and current rates of slope erosion, a surface coarse sand horizon 30–50 cm thick and a buried stone layer 30–50 cm thick would only need about 10,000–15,000 years to form.

The distinction is important because some workers have interpreted the buried stone layers that are widespread in parts of semi-arid South America as being caused by an abrupt change of climate, in which a surface lag gravel formed under arid



Figure 15.2. Stone-layer formed by termite activity, Northern Territory, Australia.



Figure 15.3. *Nasutitermes triodiae* termite mounds, Northern Territory, Australia.

conditions was subsequently buried beneath alluvial and colluvial sands laid down during a wetter climatic interval (Fairbridge and Finkl, 1984). If the stone layer is developed in situ as a result of termites removing sand, silt and clay from a deeply weathered mantle rich in vein quartz or pegmatite, as in many areas of basement complex rocks, then there is no need to invoke climatic change. This example neatly illustrates the concept of *equifinality*, or convergence of form, in which landscapes of similar form may arise from the operation of entirely different sets of processes.

15.7 Identification and interpretation of paleosols

Paleosols, or buried fossil soils, are widespread in every arid and semi-arid region of the world (Figure 15.4). Not all such soils are buried; many crop out at or near the surface, often in truncated form. Within dunes, the presence of buried soils denotes a phase of stability that is usually associated with a wetter climate and widespread vegetation cover (Singhvi et al., 2010). Similarly, the regular alternation of loess and intercalated paleosols developed on weathered loess denotes an alternation of cold, dry, windy climatic intervals in which loess is deposited downwind and warm, wet intervals that allow plant growth to thrive and soils to develop (Liu and Ding, 1998). Red, clay-rich soils and dark cracking clay soils ('vertisols') in the heart of the Sahara are likewise diagnostic of wetter times (Rognon, 1967; Williams et al., 1987). However, because soil development depends on five independent variables, it is very hard to quantify the temperature and precipitation regime under which those soils developed. A well-defined textural B-horizon may denote an abundance of wind-blown dust and an efficient leaching regime under seasonally wet conditions, or it may reflect a very long interval of moderate soil formation. In the absence of other independent means of quantifying past climatic or other environmental attributes, usually including the detection of vegetation type using pollen and phytolith analysis (Chapter 16) or carbon isotopic analysis, one is inclined to agree with Galloway (1971) that soil scientists are more likely to be consumers than producers of paleoclimatic data. With this cautionary note in mind, paleosols can still provide information that is not always obtainable by other means, as the following examples serve to illustrate.

In the far north-west of Sudan, there is a huge volcanic massif, Jebel Marra, which covers an area of 13,000 km². The highest summit on this mountain attains an elevation of 3,042 m. The massif lies at the intersection of two major tectonic lineaments that run across North Africa from south-west to north-east and from south-east to north-west, and had an estimated original volume of about 8,000 km³, compared to 3,000 km³ for Tibesti. The flora of Jebel Marra is unique and contains a mixture of plant species from the southern tropics and from the northern Mediterranean realms (Wickens, 1975a; Wickens, 1975b; Wickens, 1976a; Wickens, 1976b). Along the western and southern margins of the volcanic massif, there are cliff sections exposed where rivers have cut

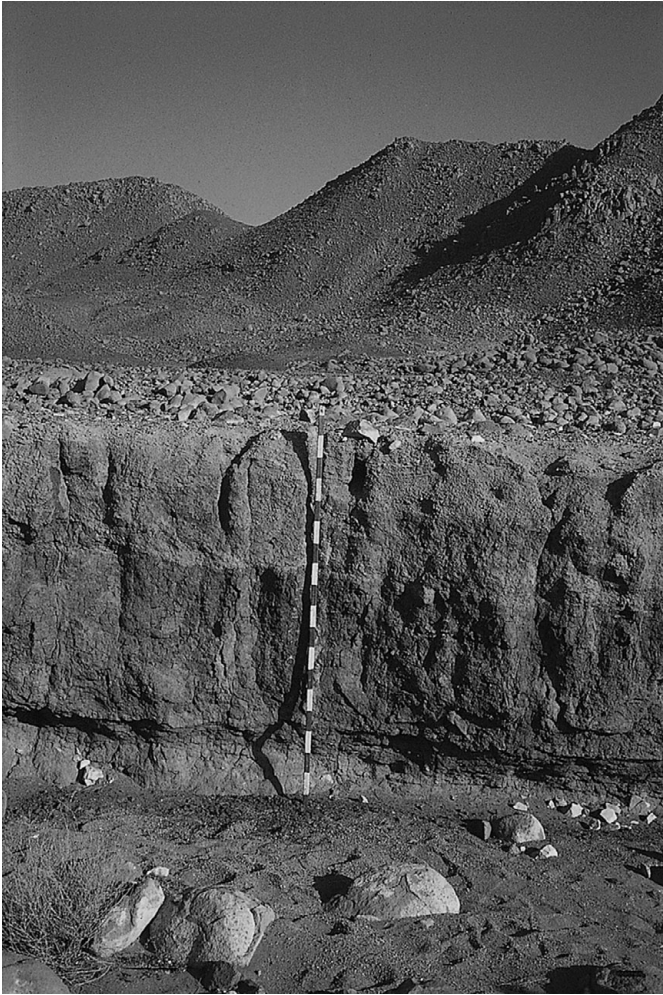


Figure 15.4. Fossil soils, Adrar Bous, south-central Sahara.

through the piedmont deposits flanking the mountain. The sections between Kas and Nyertete near the village of Umm Mari contain fossil oil palm (*Elaeis guineensis*) and *Combretum* leaf impressions embedded in reworked volcanic tuffs (Wickens, 1975b; Wickens, 1976a; Wickens, 1976b; Williams et al., 1980), as well as occasional red paleosols developed on sediments derived from the weathered Precambrian Basement Complex rocks that underlie and surround the volcanic massif (Figure 15.5). The oil palm fossils indicate the former presence of tropical rainforest, but the area is semi-arid today. Originally thought to be of Holocene age, the presence of Developed Oldowan/Early Acheulian stone tools (see Chapter 17) found in association with the fossils and paleosols indicates an age between about 1.5 and 0.8 Ma for the soils and plant fossils (Williams et al., 1980).

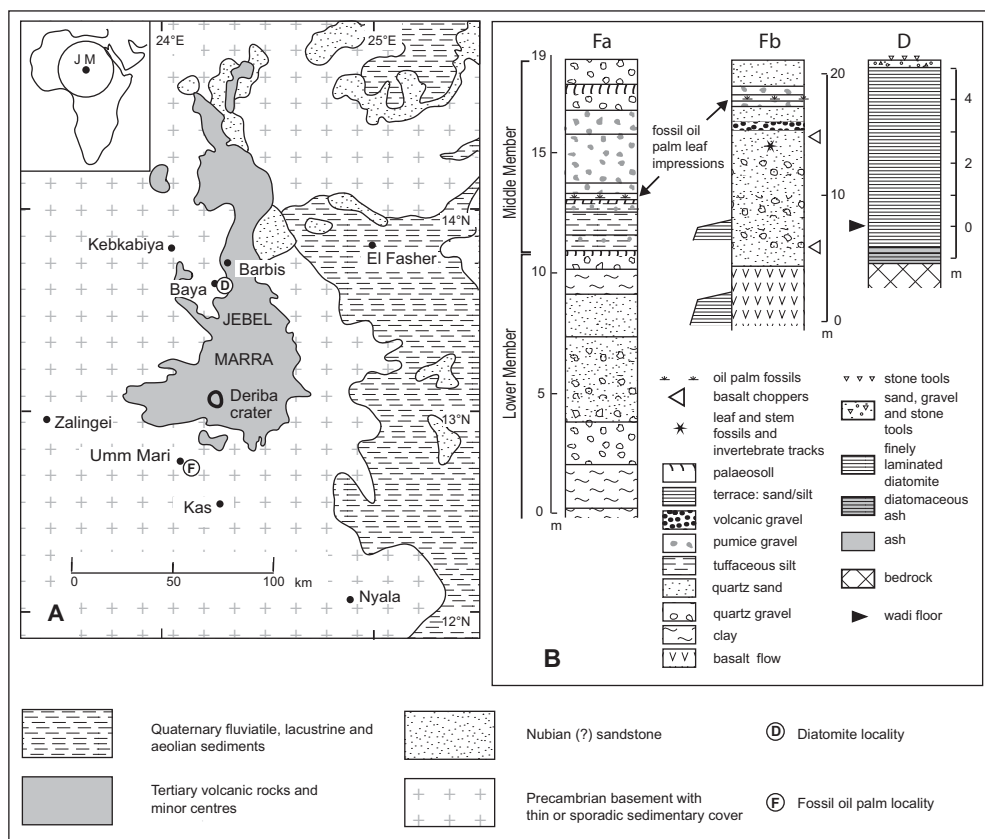


Figure 15.5. Paleosols associated with prehistoric artefacts and oil palm leaf fossils in the piedmont zone of Jebel Marra volcano, north-west Sudan. (After Philibert et al., 2010.)

In the West Kenya Rift, a series of red clay paleosols formed from volcanic ash occur in association with Middle Stone Age artefacts dated to at least 200 ka. Of interest here is the fact that both wet and dry Munsell colours on these soils are identical, which is not the case with younger soils in this region. The reason for this phenomenon is not known but may imply that the soil pedes become progressively denser over time, as a result of pressure from overlying sediments and/or as a result of the infilling of pore spaces with clay particles. The wet and dry colours of paleosols in this region thus provide a relative age sequence.

Paleosols in the Middle Awash Valley of the Afar Desert range in age from early Pliocene to late Pleistocene, and are intercalated between alluvial, lacustrine and volcanic sediments (Williams et al., 1986; WoldeGabriel et al., 2009) that contain plant microfossils (pollen, phytoliths) and vertebrate (including hominid) fossils (see Chapter 17). Analysis of the stable isotopic composition of pedogenic carbonates

within the paleosols has provided supporting evidence of the changing mosaic of habitats in which the early hominids evolved (WoldeGabriel et al., 2009).

Amit and her co-workers have investigated soil-forming processes in the hyper-arid southern Negev Desert and have used the insights gained from this work to analyse seismic hazard and earthquake frequency in the Arava Valley section of the Dead Sea Transform Fault (Amit et al., 1999). Her more recent work has made innovative use of the desert Reg soils of the southern Negev to show that this region has remained hyper-arid throughout much of the Quaternary, contrary to earlier speculations (Amit et al., 2006). Finally, she and her co-workers have demonstrated that the coarser silt particles in the primary loess deposits preserved on isolated ridge tops in the Negev were most probably derived from a proximal dune source and fashioned by wind abrasion (Crouvi et al., 2008).

Indeed, wind-blown dust plays a major role in the formation of desert soils, and it is hard to overestimate its importance (Pye, 1987; Liu et al., 1985; Liu, 1987; Williams et al., 1991a). Dust accumulates on the surface of stony deserts and especially in the crevices between individual stones. Some of this dust is then washed down into the underlying substrate, provided that the substrate is sufficiently porous. Over time, a layer of silty clay develops above the bedrock surface and becomes progressively thicker with age. If the desert dust is calcareous, as is often the case, successive cycles of leaching, solution and precipitation of dissolved carbonate will result in the formation of pedogenic carbonate within the subsoil. This layer is often protected from wind and water erosion by a desert pavement, defined as ‘armoured surfaces composed of angular or rounded fragments, usually one or two stones thick, set on or in matrices of finer material comprising varying mixtures of sand, silt or clay’ (Cooke et al., 1993, p. 68). Desert pavements can form in a variety of ways or through a combination of processes. Winnowing of finer particles from the surface by wind or running water can lead to the formation of a protective layer of surface stones sometimes called a lag gravel. Another process involves the upward displacement of stones as a result of expansion and contraction of the soil, usually caused by wetting and drying, a process that gradually moves the larger particles to the surface. The rocks forming the desert pavement will ultimately break down as a result of a variety of chemical and physical processes, including frost shattering, salt weathering and chemical attack from snails and lichens. Amit et al. (1993) have identified a series of distinct stages in the shattering of the surface gravels of desert pavements as a result of salt weathering.

Soils formed on alluvial terraces in deserts often contain a mixture of parent materials. In the central Pilbara region of north-west Australia, the bedrock consists of Precambrian ironstone that has undergone multiple cycles of weathering and erosion. The Little Sandy Desert lies some 200 km to the east – too far to be a direct source of wind-blown sand. However, during times when the prevailing easterly winds were

stronger and gustier than today, particularly during colder, drier intervals in the past 100 ka, when the dune fields and alluvial outwash areas east of the Pilbara were far more sparsely vegetated, considerable volumes of wind-blown dust were blown from the arid centre of Australia in a north-west direction out to sea (Bowler, 1973; Bowler, 1976). Wind-blown dust is as a rule very well sorted, with the grains rounded to sub-rounded and a high proportion of quartz particles. More locally derived dust can be bimodal in terms of particle size but generally better sorted than fine alluvial sediments. The Pilbara would have been in the pathway of this very wide dust plume. Dust scavenged from the atmosphere during sporadic rains would be deposited downwind of the source, and it would therefore be available for reworking by local streams. The particle size of dust is usually in the silt to clay fraction, that is, generally finer than about 15 to 20 μm , but within the longitudinal axis of the dust plume, particles as coarse as very fine sandy silt (20–60 μm) can be transported for several hundreds of kilometres downwind before deposition from suspension. Fine sand is more likely to have been transported by saltation near the base of the dust plume (Pye, 1987). It is therefore quite possible that the ephemeral streams were carrying a mixed load of both locally derived ironstone sand and gravel and very fine sand of fluvio-eolian origin that had undergone multiple cycles of reworking. The resulting weakly developed fluvisols/entisols (Tables 15.1 and 5.2) on the terrace surfaces in the Pilbara region reflect this polygenic inheritance.

Certain late Pleistocene valley-fill deposits in arid mountainous areas are derived almost entirely from reworked desert dust or loess. Examples include the Matmata Hills in Tunisia (Coudé-Gaussen et al., 1987) (Figure 15.6), the Flinders Ranges in South Australia (Williams et al., 2001; Haberlah et al., 2010a; Haberlah et al., 2010b), the Sinai Desert (Rögner et al., 2004) and the Namib (Eitel et al., 2001; Eitel et al., 2005). In all four of these regions, the fine-grained alluvial silts and clays display one or more intercalated paleosols within the main alluvial sequence, which is indicative of a brief halt to valley aggradation with sufficient time for plants to colonise the alluvial surface and for organically enriched soils to develop, a process that might take decades or centuries. In the seasonally wet Son Valley in north-central India, a vertically stacked sequence of soils has formed on late Pleistocene and early Holocene silty levee deposits, with pedogenic carbonate nodules common within each soil (Williams and Clarke, 1995).

The vertical alternation of parent material–soil–parent material in a variety of geomorphic contexts (eolian, alluvial, colluvial) prompted Butler (1959; 1967) to coin the term K-cycle for each such couplet identified in south-east Australia (Butler, 1967). In descending stratigraphic order, each couplet was labelled K_1 , K_2 , K_3 and so on. Soils were presumed to have formed during intervals of landscape stability and to have been buried during ensuing phases of landscape instability. Valiant attempts to correlate individual K-cycles were ultimately thwarted by poor chronologic control, by the realisation that many factors can contribute to local erosion and by the



Figure 15.6. Alluvial terrace composed in part of reworked loess, Matmata Hills, Tunisia.

likelihood of major gaps in the sequence. Other workers have independently come up with similar notions. Erhart (1967) used the term *rhexistasis* for sudden intervals of landscape instability during which previously stable soil mantles, formed during times of landscape biological equilibrium (or *biostasis*), were abruptly eroded. Causes initiating such disequilibrium could include forest destruction by humans, fire or climate change. As with Butler's K-cycle model, independent verification of any climatic inferences that are drawn from this model is essential, as is rigorous dating of each phase of landscape stability and instability. It is unlikely that these models will contribute much of value to unravelling past climatic events in any detail.

15.8 Duricrusts in deserts and desert margins

A *duricrust* is simply a hard crust. Four main types of duricrust have been identified in deserts and along their margins: *ferricrete*, *silcrete*, *calcrete* and *gypcrete*. Lamplugh (1902) coined the first three terms. He defined ferricrete as a conglomerate consisting of surficial sand and gravel cemented into a hard mass by iron oxide derived from the oxidation of percolating solutions of iron salts. Silcrete he defined as a conglomerate consisting of surficial sand and gravel cemented into a hard mass by silica. Calcrete was a conglomerate cemented into a hard mass by calcium carbonate precipitated

from solution and redeposited either through the agency of infiltrating waters or from the escape of carbon dioxide from vadose water (Lamplugh, 1902; Bates and Jackson, 1987, p. 94). It follows from these definitions that ferricrete, silcrete and calcrete are secondary features formed by the precipitation of allochthonous inputs of iron, silica or calcium carbonate within coarse-grained alluvium. In the case of certain iron- and silica-rich formations, this is not always true. Gypcrete, as the name implies, is a hard gypsum crust. Ferricrete in Lamplugh's sense of the term is broadly equivalent to certain forms of detrital laterite, discussed in [Section 15.8.2](#).

Gypsum crusts are less widespread than calcrete, silcrete or ferricrete are because of the greater ease with which gypsum can be dissolved or removed by deflation. Gypcrete tends to be confined to playa and sebkha margins in hyper-arid regions, as in the northern Sahara, the coastal regions around the Persian Gulf, the Lake Eyre Basin in central Australia and a number of the saline lakes in the more arid regions of South America (Watson, 1983, [fig. 5.1](#)). As a broad generalisation, calcretes and silcretes tend to occur in the semi-arid to arid regions of Africa and Australia in particular, while ferricretes occupy the more humid sectors of the desert margins (Mabbutt, 1977). The value of silcretes and ferricretes as indicators of past climates is limited by the difficulty of obtaining precise and reliable ages for these formations (Bourman, 1993; Twidale and Bourne, 1998) and, in the case of silcrete especially, by the lack of convincing modern analogues.

15.8.1 Pedogenic, biogenic and groundwater calcretes

Calcretes have been widely investigated in many parts of the semi-arid world (Goudie, 1983; Lekach et al., 1998; Khadkikar et al., 2000; Amit et al., 2007; Amit et al., 2010; Singhvi et al., 2010), not least because they are often hosts for economically important minerals such as uranium, copper and gold. Calcretes have attracted considerable recent interest from the mineral exploration community, most notably in Australia, where a Cooperative Research Centre for Landscape Environments and Mineral Exploration involving the mining industry, universities and government agencies has been pioneering new methods of detecting such minerals as uranium, copper and gold in calcrete deposits formed within desert dunes, fossil river channels and desert lake margins (Lintern, 2001; Keeling, 2004; Schmidt Mumm and Reith, 2004; Wittwer et al., 2004). Uranium-bearing calcretes in Mauritania and Namibia have also aroused the interest of exploration geologists but have so far been less intensively investigated than their Australian counterparts.

Calcareous soils are actively forming today on a variety of rock types in areas with a mean annual rainfall of 200–500 mm, although they may occur in wetter areas where the underlying parent rock is limestone, dolomite or eolianite. A series of stages has been identified in the development of calcretes in the drier regions of the United States, north-west and southern Africa, north-west India, Israel, peninsular Arabia,

and central and southern Australia (Goudie, 1983). An early stage is the precipitation of soft powdery aggregates of calcium carbonate (CaCO_3) within the profile. The source of such carbonate may be groundwater, run-off, wind-blown dust or the parent material. As time progresses, the soft aggregates harden into irregular nodules or spherical concretions that may eventually occupy more than half of the soil horizon in which they occur. Later still, the hard nodules become cemented into a single massive unit (Khadkikar et al., 2000). Solution and reprecipitation of carbonate can occur at any stage. The chemical equations governing solution and precipitation were given in Chapter 14 (Equations 14.1, 14.2 and 14.3).

Water (H_2O) combined with carbon dioxide (CO_2) forms carbonic acid (H_2CO_3). Carbonic acid dissolves calcium carbonate (CaCO_3) to form calcium bicarbonate in solution [$\text{Ca}(\text{HCO}_3)_2$]. An increase in temperature or a decrease in pressure cause the dissolved carbon dioxide to come out of solution, and calcium carbonate is then precipitated. Biological activity can also assist in the precipitation of calcium carbonate. Carbonate coatings can be seen forming today around exposed pine tree roots growing on sand dunes in Algeria, as well as around acacia tree roots exposed along the banks of the lower Blue Nile in central Sudan and around River Red Gum (*Eucalyptus camaldulensis*) roots in the arid Flinders Ranges of South Australia.

Many calcretes are neither pedogenic nor biogenic in origin but have formed as a result of the subsurface lateral movement of groundwater (Ruellan, 1968). Such calcretes are often finely laminated and often occur at shallow depth along the margin of former lakes and ponds, or else crop out along the scalded margins of fixed dunes (Williams, 1968a).

Vertically stacked beds of pedogenic carbonate nodules are found within Pleistocene river levee deposits along the seasonally arid Son Valley in north-central India, as well as in alluvial fan deposits in semi-arid west-central New South Wales (Williams et al., 1991a). Both examples are indicative of successive phases of alluvial deposition interspersed with dry phases during which there is seasonal leaching and precipitation of calcium carbonate below the base of the soil-wetting zone.

One advantage of using calcretes to reconstruct past changes in climate rests on the relative ease with which they can be dated. Initial dating efforts were confined to the use of radiocarbon, but problems arose once the 35,000-year upper limit of conventional radiocarbon (^{14}C) dating methods was attained. At this point, even slight amounts of contamination by both younger and still radioactive ^{14}C , perhaps derived from plant rootlets or from old, inert carbon present in groundwater, could give spurious results, as discussed in Chapter 6. This limitation was in part overcome through the use of uranium-series dating which soon demonstrated that many samples from Saharan lake calcretes deemed to be very late Pleistocene in age were in fact of last interglacial age (Causse et al., 1988; Fontes et al., 1992). The method can yield reliable ages provided there are no external gains or losses of uranium, that is, when we are dealing with a closed geochemical system. Furthermore, inputs of

detrital thorium from wind-blown dust, for example, can again lead to inaccuracies in dating.

One especially useful tool in determining the likely provenance of the initial calcrete parent material involves the use of stable strontium (Sr) isotope analysis. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio in the calcrete reflects the geology of the initial source of the calcium in calcium carbonate and does not alter over time. Dart et al. (2004; 2007) have used this technique to demonstrate that valley-floor calcretes located more than 400 km from the present coast in arid and semi-arid inland Australia, once thought to have formed by in situ weathering of local rocks, were in fact derived primarily from wind-blown dust blown from coastal eolianites and from the exposed continental shelf during times of lower sea level.

15.8.2 *Equivocal climatic significance of laterite and lateritic soils*

Although in popular usage laterites are generally associated with the hot, wet tropics or the seasonally wet tropics, laterite remnants occur sporadically in many deserts and semi-deserts. The horizontally bedded Mesozoic sandstone hills conspicuous in the deserts of Africa, Arabia, India and Australia have been preserved from erosion by a resistant caprock of ferricrete, or laterite, often silicified. If we can narrow down the conditions under which such ferruginous duricrusts formed and determine their geological ages, we will be able to identify what types of environments (and, perhaps, climates) were once prevalent in those presently arid regions, although such data are very unlikely ever to yield high-resolution information.

Many authors have considered laterite and lateritic soils to be good indicators of a hot, wet tropical climate (Cooke, 1958; Flint, 1959b). As a generalisation, laterites are rich in the hydrated sesquioxides of iron and/or aluminium, and are depleted in bases, alkali earths and silica (Harrison, 1933; Sivarajasingham et al., 1962). Because they are thought to require a tropical climate for their formation, so the argument runs, then in those localities where they occur outside the tropics, the climate must once have been tropical. There are, however, a number of problems with this conclusion. Paton and Williams (1972) have provided a detailed historical review of how the term laterite has been so changed over time that it now applies to a whole variety of materials and has little diagnostic value. Originally laterite was defined by Buchanan (1807, pp. 440–441) as a clay located on the uplifted coastal plain of Malabar in India which, on exposure to the atmosphere, hardened to a brick-like consistency (hence ‘laterite’, from the Latin word *lateritis*, or ‘brickstone’). This definition excluded high-level ironstone cappings, which many later workers regarded as laterite (McFarlane, 1976; McFarlane, 1983). Subsequent investigators drew a distinction between autochthonous (primary) and allochthonous (secondary) forms of laterite, that is, between profiles in which there was a relative enrichment in iron as a result of weathering processes that led to leaching and removal of bases and silica and those in which there was

an absolute enrichment in iron from external sources (Paton and Williams, 1972; McFarlane, 1976; McFarlane, 1983).

Lithology plays an important role in weathering, and its influence may outweigh that of climate. For example, Harrison (1910; 1933) noted that primary laterite failed to develop on granite bedrock in British Guiana, but did form on basic rocks in that region. The notion that laterite invariably demanded a tropical climate was effectively refuted by Goldschmidt (1928) who found it actively forming on labradorite in Norway, by Robinson (1932) who observed it forming on steep slopes in North Wales and by Crompton (1960) who found it forming on basic to intermediate lavas on very steep slopes under high rainfall in the English Lake District. In all three cases, an efficient leaching regime was characteristic. Milne (1938; 1947) reached a somewhat similar conclusion from his soil mapping in East Africa but considered that laterites were more widespread in tropical regions because higher soil temperatures speeded up the processes of chemical weathering. In fact, measurements of dissolved river loads have shown that rates of chemical erosion are no faster in the tropics than they are in temperate and periglacial regions (Livingstone, 1963; Davis, 1964; Gibbs, 1967; Douglas, 1969).

Extensive regions in the Sahara and inland Australia consist of horizontally bedded Mesozoic sandstones capped by resistant beds that are rich in iron (Bourman, 1995; Twidale and Bourne, 1998). These caprocks have been variously termed 'ferricrete', 'ferruginous duricrusts', 'cuirasses ferrugineuses' and 'ironstone'. Some authors have claimed that these ferruginous duricrusts are genetically linked to the final stages of peneplanation and therefore imply prolonged tectonic stability (Woolnough, 1927). Despite this questionable assumption, the presence of laterite surfaces at varying elevations has been used as a guide in reconstructing the Cenozoic tectonic and denudational history of large tracts of southern Africa and northern Australia (Wright, 1963; Hays, 1967; Maud, 1968; Twidale and Bourne, 1998). In addition, Glassford and Semeniuk (1995) have proposed that the 'lateritic sandplains' of south-west Australia are in fact allochthonous eolian deposits and not the result of deep weathering and laterite formation in situ.

The use of laterite to correlate erosion surfaces assumes that the laterite is the same age as the surface it overlies, which may not be true (Bourman, 1993; Bourman, 1995). Williams (1969b) mapped the distribution of morphologically distinct types of laterite on a tract of seasonally wet tropical Australia 40,000 km² in area and found that in certain sites laterites were actively forming, in others they were stable and in others they were disintegrating. In addition, profiles once regarded as matching the so-called standard monogenetic lateritic profile as defined by Walther (1915; 1916) (consisting from the base up of a pallid zone, a mottled zone and a ferruginous zone) were seen, on close examination, to consist of a younger, transported layer of ironstone gravel lying unconformably above a slightly mottled and kaolinised pallid zone (Williams, 1969b).

In conclusion, ‘the fact that the single term laterite has been applied to materials as diverse and genetically distinct as iron-cemented colluvial rubble, weathered basalt, mottled clays, and kaolinized igneous rocks has caused much unnecessary confusion’ (Paton and Williams, 1972, p. 55). These authors go on to say that ‘perhaps the greatest misconception relating to laterite is the notion that laterization (an ill-defined and ill-understood complex of weathering processes) demands a tropical climate’ (op. cit., p. 55). To do so is to ignore the influence of the other, equally important soil-forming factors of parent material, topography, drainage and time. Nevertheless, the presence of laterite or lateritic soils within presently arid regions is clearly indicative of a previously efficient weathering and leaching regime under a climate that must have been substantially wetter (Bourman, 1995; Twidale and Bourne, 1998). The Eocene deep weathering profiles of the south-central Sahara are a case in point (Greigert and Pougnet, 1967), as are the Mesozoic and Cenozoic laterites of central Australia (Mabbutt, 1965b).

15.9 Equivocal paleoclimatic significance of silcrete

Summerfield (1983) has provided a comprehensive review of the distribution and possible origins of silcretes in the Namib Desert. He concluded that silcretes enriched in titanium occurred today in areas of more humid climate and that those that were titanium-poor had apparently formed in arid to semi-arid environments. However, because the underlying bedrock geology also appears to exert a strong control over the distribution of silcrete, at least in arid inland Australia (Mabbutt, 1965b), and because the mode or modes of formation of silcrete remain poorly understood, the presence of silcrete is not particularly useful as an indicator of past climate. Reprecipitation of quartz derived from sandstone bedrock during weathering may also play a role in silicon cycling and the eventual formation of silcrete units (Basile-Doelsch et al., 2005).

15.10 Conclusion

Any landscape that supports plant life comprises a surface soil mantle. Soils are formed as a result of five main factors: parent material, topography, climate, biological activity and time. In young soils, the first two of these factors (parent material and topography) exert the dominant influence on soil morphology and soil chemistry, with immature soils formed on desert dunes being a good example. As time progresses, soil climate and biological processes within the soil will exert increasingly important control over the soil’s physical and chemical properties. The classic subdivision of soil profiles into A-, B- and C-horizons is not always a useful basis for classifying soils, because quite different types of soil profile can develop as a result of the sorting



Figure 15.7. Contorted Cenozoic salt lake sediments, Negev Desert, Israel.

processes engendered by insects, such as ants or termites, and of the subsequent erosional reworking of material brought to the surface by insects and other soil fauna.

The presence of relict soils, or paleosols, in areas that are now arid indicates that conditions were previously sufficiently humid for plant growth to occur there. However, it is not always possible to distinguish between the effects of short intervals of intense humidity and rapid soil development and longer intervals of lesser humidity and slower soil development, given that the end results may be very similar.

Stratified sequences of buried soils within loess deposits, river and lake sediments, and volcanic ash beds can provide potentially useful information about past phases of landscape stability during which the climate was wetter, widespread plant growth was possible and soils could form.

The soils that develop under extreme aridity reflect the influence of sporadic rainfall events and the precipitation of gypsum, halite and/or calcium carbonate within the soil profile. Progressive accumulation of gypsum or calcium carbonate within the soil profile can ultimately lead to the formation of hard beds or crusts of calcrete or gypcrete and no further soil development. In the Negev Desert, laminated sediments composed of alternating layers of gypsum and silt have been contorted by recent earth movements (Figure 15.7).

In the seasonally wet tropics, the mobilisation of silica and bases and the reprecipitation of iron or aluminium results in the formation of various forms of laterite profile. Apart from indicating an efficient leaching regime, laterite soils and crusts have little paleoclimatic value. Silcretes formed as a result of near-surface precipitation of dissolved silica appear to form most extensively in areas that are less arid than those where calcretes occur but are more arid than areas where laterites occur. Overall, desert soils, paleosols and duricrusts do not, in general, provide much in the way of precise paleoclimatic data.

16

Plant and animal fossils in deserts

The discovery of fossil remains has been a very slow
and fortuitous process.

Charles Darwin (1808–1882)

The Descent of Man (1871)

16.1 Introduction

Fossils have always played a major role in geology. For example, the geological time scale was established on the basis of the presence or absence of certain fossils, with particular emphasis on past extinctions, such as the disappearance of dinosaurs at the end of the Mesozoic era 65.5 million years ago, as well as the first appearances of certain distinctive plants and animals. The fossil fauna of our best-known deserts, such as the Sahara and Gobi, bear witness to a vanished era when dinosaurs roamed these once green lands. In addition to the fossilised bones of the former vertebrate fauna, silicified tree trunks scattered across many present-day deserts point to a time when they were once forested.

In geologically more recent times, only a few thousand years ago, small herds of antelope, Cape buffalo, and occasional elephants and giraffes occupied the once wooded grasslands of the Sahara (Jousse, 2004), whose rivers and lakes supported an aquatic fauna of fishes (including large Nile perch), turtles, hippos and crocodiles (Vernet, 1995). However, caution is needed here to not extrapolate too boldly from the sporadic fossil remains. As Gautier et al. (1994) have reminded us: ‘One elephant doesn’t make a savanna’.

Prehistoric rock art is an indirect form of fossil evidence. Scattered across the Sahara, there are magnificent rock art galleries with depictions of the savanna fauna engraved or painted wherever suitable rock outcrops offered fresh, smooth surfaces. The older paintings focus on such animals as giraffes, elephants, gerenuks and ostriches, while the later Neolithic images show scenes of cattle herding, including

moving camp and fishing from papyrus canoes (Muzzolini, 1995; Coulson and Campbell, 2001). Prehistoric archaeological sites and the associated fossils of plants and animals provide tangible evidence that the rock paintings and rock engravings often represent day-to-day events in the lives of these prehistoric desert peoples. The plant and animal fossils found in the lake and river sediments and in the relict soils preserved in the deserts provide additional evidence with which to reconstruct past environments and climates in these now arid regions.

The marine microfossil record off the coast of desert regions is often even more useful than the terrestrial fossil record of past environments because it is generally more continuous, is amenable to analysis of the stable isotopic composition of the microfossils, notably foraminifera, and can usually be accurately dated. The aim of this chapter is to consider the scope and limitations of using plant and animal fossils to reconstruct climatic change in deserts. The subject is vast, but happily many excellent specialist accounts are readily available relating to the use of fossil plants and animals in reconstructing past environments on land and sea (Hill, 1994b; Vrba et al., 1995; Smol et al., 2001a; Smol et al., 2001b; McGowran, 2005).

16.2 Desert refugia and disjunct distributions of plants and animals

In a comprehensive and still useful review of prehistoric environments in the Sahara, the great Saharan scholar Théodore Monod noted that some elements of the earlier savanna fauna have survived, albeit rather precariously, until the present (Monod, 1963). In sheltered valleys in the rugged Air Massif in Niger, near the southern margin of the Sahara, there are remnant populations of baboons (*Papio anubis*) and patas monkeys (*Cercopithecus patas*), which must have reached these desert mountains during times of wetter climate when the West African savanna woodland was more extensive than it is today (Monod, 1963). They probably made use of riparian forest corridors that grew along once permanent rivers flowing south from the mountains. The Awash River that flows from the Ethiopian Highlands down into the Afar Desert is a possible modern analogue, in that its banks and floodplain are thickly wooded and support a modest primate fauna. The dwarf crocodiles that lived in some of the permanent waterholes in the Tibesti Mountains of the south-central Sahara until the 1950s but have since been hunted to extinction were part of this climatic legacy (Lambert, 1984).

In the same volume in which Monod's overview appeared, Moreau (1963) published an important paper relating to the montane avifauna of Africa. He argued that montane evergreen forest bird populations that were hitherto in contact became isolated during geologically recent times of colder, drier climate, when the forests themselves became isolated on particular mountains. A comparable study of disjunct populations, this time of Amazonian *Heliconius* butterflies, concluded that fragmentation of the Amazon rainforest during drier climatic intervals led to the isolation of the butterflies and is

reflected in subtle differences between populations (Brown et al., 1974). Whitmore and Prance (1987) have since identified overlapping centres of endemism of butterflies, plants and birds in South American rainforests, which seem to reflect the existence of refugia during times of drier and/or colder climate.

Based on pollen analysis from a somewhat limited number of sites, Anhuif et al. (2006) concluded that the area now covered by humid rainforest in the Amazon was probably reduced by around 54 per cent as a result of a 20–40 per cent decrease in precipitation accompanied by a temperature drop of 4.5–5°C during the LGM. The presence of now vegetated sand dunes within the Amazon Basin is also indicative of previously drier conditions that were conducive to the formation of source-bordering dunes. This was also a time when the area of the rainforest in the Congo Basin possibly shrank by 84 per cent as a result of a 30–40 per cent reduction in rainfall and a 5°C drop in temperature during the LGM (Anhuif et al., 2006), as inferred qualitatively by Moreau forty years earlier (Moreau, 1963). As we shall see in Chapters 18 and 19, elsewhere in many parts of tropical Africa and Asia, the deserts expanded, woodland areas contracted and lake levels fell during the LGM. However, there was considerable regional diversity during the LGM in Africa (Gasse et al., 2008), and some workers disagree with the LGM glacial aridity scenario for the Amazon Basin (Colinvaux et al., 1996; Colinvaux et al., 2000; Colinvaux, 2001).

In Australia, Byrne et al. (2008) have sought to integrate evidence from phylogenetics, phylogeography and paleoenvironmental studies in order to reconstruct when and how the present-day Australia arid zone biota arose. They concluded that aridity first became evident in the plant record in the mid-Miocene some 15 million years ago. Landforms consistent with full aridity (such as dunes and stony gibber plains) appear in central Australia between 4 and 1 million years ago (Fujioka et al., 2005; Fujioka et al., 2009). Dated molecular phylogenies indicate that some large, vagile taxa show patterns of recent expansion and migration throughout the arid zone, while other taxa appear to have persisted in multiple localised refugia during cold, dry glacial times (Byrne, 2008a; Byrne, 2008b; Byrne et al., 2008). No similar studies have yet been published for other major desert regions, such as the Sahara, the Gobi or the deserts of central Asia, but this seems a potentially fruitful approach, drawing on insights from both molecular ecology and the earth sciences. Indeed, Hewitt (2000) argued that the use of DNA technology can show how different organisms have responded to the climatic vicissitudes of the Quaternary ice ages and concluded that ‘the present genetic structure of populations, species and communities’ has been mainly formed by the environmental changes associated with Quaternary glaciations.

16.3 Vertebrate fossils: life and death assemblages

With rare exceptions, vertebrate remains are at best sporadic in most of the arid zone. When they do occur, they tend to be found in association with prehistoric occupation



Figure 16.1. Molar of Pliocene *Elephas recki* used to build stone enclosure for baby goats, Afar Desert, Ethiopia.

sites (Jousse, 2004) or with former lakes (Wendorf et al., 1993; Gautier et al., 1994; Vernet, 1995; Tillet, 1997). Fossil bones found in river sediments have generally been reworked and suffer from the results of differential transport and differential preservation. In the case of smaller mammals, the bones are often broken up and highly weathered, making them difficult to date. Exceptions to the general rarity of vertebrate fossil remains in deserts are the East African and Afar rifts, which have yielded an abundance of well-preserved bones and teeth (Vrba et al., 1995; de Heinzelin et al., 2000) (Figures 16.1 and 16.2), and the limestone caves of semi-arid southern Africa (Brain, 1981b) and of semi-arid South Australia (Macken et al., 2012). In all of these cases, caution is needed when seeking to infer past habitats from the fossil remains, because the remains will reflect location at the time of death, together with any disturbance from scavenging carnivores and vultures, followed by the possible effects of mass movement and run-off. An entire branch of vertebrate palaeontology, taphonomy, is devoted to unravelling these effects. Given that the rate of speciation of large vertebrates is relatively slow (amounting to about 0.5 Ma for large bovids), an independent means of dating the fossil assemblage is essential if they are to provide useful information about past ecosystems and climates. One vertebrate that has been the focus of unparalleled study is the hominid family, which includes both modern humans (*Homo sapiens*) and our prehistoric ancestors, the *Australopithecines*, detailed in the next chapter (Chapter 17).



Figure 16.2. Pliocene pig mandible, Afar Desert, Ethiopia.

Once the vertebrate fossil has been identified, preferably to species level but if not at least to genus level, the next step is to define the ecosystem within which such animals occur (Jousse, 2004). Usually a number of different animal genera are represented within any given fossil assemblage, so it becomes possible to define the former habitats more precisely. If the animal concerned has no living counterparts and is a herbivore, it may be necessary to rely on tooth wear patterns and on the stable isotopic composition of bones and teeth in order to reconstruct probable diet (van der Merwe, 1982; Quade et al., 1989; Ayliffe and Chivas, 1990; Cerling et al., 1991; Morgan et al., 1994; WoldeGabriel et al. 1994; WoldeGabriel et al., 2001; WoldeGabriel et al., 2009; Cerling et al., 2010). Ideally, for such studies to be convincing, detailed isotopic analysis of modern herbivore teeth are necessary, including close attention to any seasonal changes in diet that may reflect responses of the plant cover to seasonal changes in precipitation sources (Brookman and Ambrose, 2012; Brookman and Ambrose, 2013). Such studies are still in their infancy.

P. deMenocal (2004) compared the African faunal record spanning the last 5 million years with the marine evidence of Pliocene-Pleistocene wetter and drier phases linked to orbital variations (see Chapter 6). He found evidence of step-like (± 0.2 Ma) changes in aridity and climatic variability at around 2.8, 1.7 and 1.0 Ma, all of which coincided with the onset and intensification of high-latitude glacial cycles. The African faunal evidence indicated more open habitats at 2.9–2.4 Ma and after 1.8 Ma, although

there were still significant gaps in the faunal record. He further concluded that times of step-function change in climate and habitat coincided with changes in the hominid record, as well as with the emergence of *Homo* at around 1.8 Ma, reviewed in the following chapter ([Chapter 17](#)).

In contrast to the coarse resolution faunal analysis of deMenocal (2004), Macken et al. (2012) were able to provide a much finer resolution analysis for the Late Pleistocene vertebrate fossils preserved in Victoria Fossil Cave at Naracoorte in semi-arid South Australia. They found that although certain species reacted to changes from warm moist to cooler drier regional climatic conditions in much the same way as they had in the past, others showed a more complex response that was variable and individual through time. They concluded that large fossil samples were needed in order to demonstrate how particular species had responded to past climatic changes.

Ambrose et al. (2007a) provide a comprehensive analysis of the late Miocene fossil fauna at the site of Lemudong'o in semi-arid southern Kenya. They were able to demonstrate changes in habitat at a lakeshore site within a mosaic of riparian forests, open woodlands and wooded grasslands. At another site, they were also able to show that an accumulation of carcasses demonstrated significant carnivore damage on the bones, probably caused by several avian and small mammalian carnivores. As a prelude to the analysis of the fossil fauna, Ambrose et al. (2003; 2007b) carried out a thorough study of the geology, geochemistry and stratigraphy of the Lemudong'o Formation within which the fossils were found, and dated certain widespread tephra marker beds. Such studies are a vital prerequisite to the analysis of any fossil vertebrate assemblage, because they can show whether disturbance by running water, mass movement or tectonic activity needs to be taken into account.

An earlier study of Miocene habitats in East Africa suggested that the progressive fragmentation of a relatively homogeneous lowland tropical rainforest as a result of rifting and climatic change led to dispersal out of the forest by some species, extinction of others and retreat to the forest remnants by others (Malone, 1987). The dispersal of Miocene hominoids (see [Chapter 17](#)) was suggested to be one such response to habitat change, as was the emergence of hominid bipedalism.

16.4 Invertebrate fossils

16.4.1 *Marine foraminifera*

Foraminifera are single-celled and mostly marine planktonic animals with a moderately resistant shell of calcium carbonate. Based on the first and last appearances of particular species of foraminifera, marine geologists have been able to provide a detailed time scale for the entire Cenozoic and to identify global and regional warming and cooling events (Zachos et al., 2001; McGowran et al., 2004; McGowran, 2005). [Figure 3.4](#) in [Chapter 3](#) shows some of the major global environmental changes

in the last 65 million years, many of which were reconstructed from changes in marine planktonic assemblages, especially the foraminifera. Another great advantage of using foraminifera is that they lend themselves to detailed analysis of changes in the isotopic composition of their calcareous shells. This has allowed subdivision of the Quaternary Period, identification of individual glacial-interglacial cycles and the erection of a well-dated marine isotope stratigraphy (Shackleton, 1967; Shackleton, 1977; Shackleton, 1987; Lisiecki and Raymo, 2005; Lisiecki and Raymo, 2007; Raymo and Huybers, 2008), which was discussed in Chapter 6.

As with vertebrate fossils, the first step is to establish very precisely the type of environment in which the present-day foraminifera species of interest are living, including the optimum habitats as well as the limits. For this, a systematic sampling programme of ocean water temperature, depth, chemistry and nutrient levels is a prerequisite for allowing the establishment of 'transfer functions' that correlate species with particular attributes of their habitats. Such transfer functions may then be applied to fossil assemblages in order to reconstruct past changes in, for example, sea surface temperature at particular times in the past. Two problems often arise when using this approach. One is that the fossil assemblage may have undergone the partial dissolution of certain species and so will not be fully representative of complete modern assemblages. The solution is to concentrate on the more robust species least prone to differential losses from solution. A second and more difficult problem occurs when the fossil assemblage has no modern counterpart. Here again, a partial solution is to focus on the particular species which do have modern equivalents.

Only a very limited part of the work on marine microfossils has been carried out in the immediate vicinity of deserts (Sarnthein et al., 1982; Leroy and Dupont, 1994; Leroy and Dupont, 1997; van der Kaars and De Deckker, 2002; van der Kaars et al., 2006). However, perhaps one of the most useful aspects of marine microfossil studies is that they provide a global context within which to place more local studies. One important aspect of this idea concerns attempts to reconstruct global patterns of sea surface temperature during the Last Glacial Maximum (Barrows and Juggins, 2005), of which the Climate Long-Range Investigation, Mapping and Prediction (CLIMAP) project is an illustrious early example (McIntyre et al., 1976; McIntyre, 1981). However, probably the greatest contribution of marine micropaleontology to studies of climatic change in deserts is in its providing of a precise time scale of global temperature and ice volume changes spanning the entire Cenozoic era (Zachos et al., 2001; McGowran, 2005).

16.4.2 Ostracods, cladocera and chironomids

Ostracods are tiny crustaceans up to about 1 mm in size with hinged or bivalve shells made of low Mg-calcite. They shed their shells as they grow, and the shells accumulate on the lake or swamp floor and become buried in lake or swamp sediments. Some species of ostracod are tolerant of moderately high levels of water salinity, while others

are not, and some can swim in shallow seasonal ponds and lay drought-resistant eggs. It is therefore possible to use the proportions of different species of known salinity tolerance to estimate past changes in lake water salinity (De Deckker, 1988; Holmes, 2001; Zhai et al., 2011). Likewise, some species prefer colder water and others warmer water, so estimates may be made about past changes in water temperature. All of this presupposes that particular species have retained the same levels of water salinity and temperature tolerance through time, which may not always have been the case.

An alternative approach to using present-day ecological tolerances to reconstruct past changes is to analyse the stable isotopic composition of the calcareous shells, notably the stable carbon and oxygen values. As with marine foraminifera, high relative concentrations of the heavier isotope of oxygen (^{18}O) denote evaporative conditions in the lake water, while low concentrations denote fresh water and relatively low rates of evaporative loss. Changes in the carbon isotope ratios can be used to infer changes in biological productivity and carbon cycling within the lake (Ito, 2001).

An additional method of analysis pioneered in the early 1980s involves measuring changes in the trace element geochemistry of ostracod shells within a given lake core. In particular, the Sr/Ca ratio in ostracod shells provides a measure of lake water salinity, and the Mg/Ca ratio provides a measure of lake water temperature (Chivas et al., 1986a; Chivas et al., 1986b; Chivas et al., 1986c; Zhai et al., 2011). However, considerable finesse is needed when interpreting the results of such analyses, because many other factors besides temperature and salinity are at play in determining ostracod shell geochemistry (Ito et al., 2003; Ito and Forester, 2009).

Cladocera are another order of crustaceans, and they are better known as water fleas. There are more than 600 species of cladocera, most of which live in fresh water, although some live in the oceans. The basic principles involved in using cladocera fossils preserved in lake and swamp sediments to reconstruct past changes in lake temperature and salinity are the same as those used with marine foraminifera. Transfer functions are established relating species assemblages to their freshwater habitats, and these transfer functions are then applied to fossil assemblages in order to reconstruct past changes in lake habitat (Korhola and Rautio, 2001). As a general rule, a number of other fossils should be used in addition to cladocera, so that a more complete picture of past environments can be constructed.

Chironomids are a family of non-biting midges. There are estimated to be more than 10,000 species worldwide, occupying a range of aquatic and semi-aquatic habitats during the larval stage. The larvae are very sensitive to lake water temperature and chemistry (Walker, 2001). Transfer functions relating water temperature and chemistry to assemblages of chironomid larvae have been successfully developed for North America, Europe, New Zealand and Tasmania but not thus far for desert areas. Work is now underway to use species assemblages and the stable oxygen isotope composition of chironomid teeth to develop transfer functions for temperature and perhaps salinity

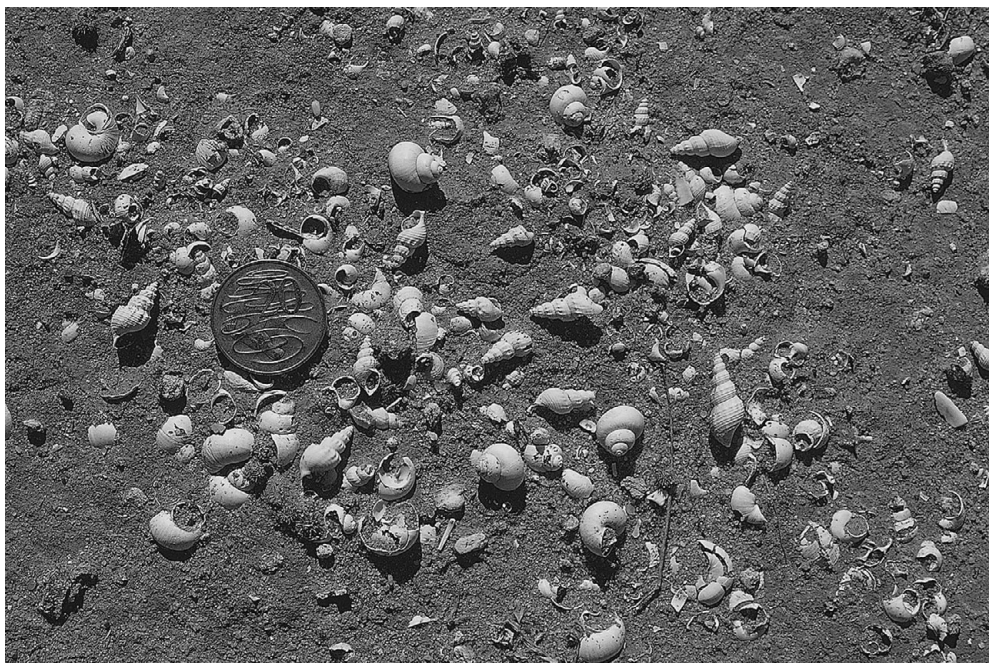


Figure 16.3. Surface shells on edge of Holocene Lake Boolaboolka, lower Darling Basin, Australia. (Photo: Don Adamson.)

in the semi-arid Snowy Mountains of south-east Australia (Jie Chang, pers. comm., December 2012).

16.4.3 *Non-marine mollusca*

One advantage of using fossil aquatic mollusca to reconstruct past environmental changes stems from their relative abundance and good state of preservation in fluvial and lacustrine sediments (Miller and Tevesz, 2001) (Figure 16.3). For example, Williamson (1982) was able to carry out a detailed analysis of molluscan biostratigraphy of the hominid-bearing deposits at Koobi Fora in north Kenya and found serious stratigraphic miscorrelations between several local sections where the correlation was based solely on fossil vertebrate evidence. In north-west Cape Province, Kent and Gribnitz (1985) used freshwater shell deposits associated with higher lake levels as additional evidence of a widespread wet phase during the late Pleistocene, when rainfall was higher, temperatures were lower and evaporation was much reduced.

African mollusca have been studied in great detail (Brown, 1980; Van Damme, 1984; Brown, 1994), not least because two genera, *Bulinus* and *Biomphalaria*, are vectors of the water-borne parasitic disease schistosomiasis. These studies have provided

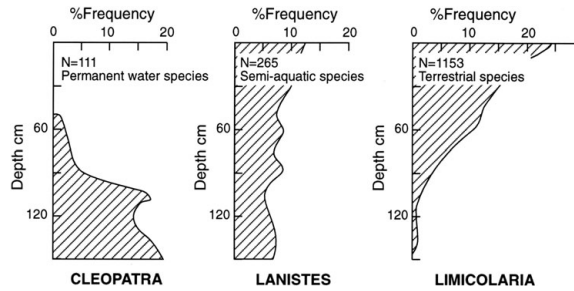


Figure 16.4. Changes in the proportions of aquatic, semi-aquatic and terrestrial snails in Gezira clay, lower Blue and White Nile valleys, central Sudan, between 15 and 5 ka. (After Williams et al., 1982, fig. 9.8A.)

valuable information about the ecological requirements of particular species, which has been applied to reconstruct the paleoecology of the Holocene lakes at Adrar Bous in the central Sahara (Williams et al., 1987) and at Erkowit mist oasis in the Red Sea Hills of eastern Sudan, which enjoyed a less arid climate 2,000 years ago (Mawson and Williams, 1984). Indeed, the combined mollusc and ostracod faunas show that now ephemeral stream channels in the Red Sea Hills were previously perennial and flowed through swampy meadows, which also, according to the contemporary Sicilian historian Diodorus, provided a refuge for local cattle rustlers (Mawson and Williams, 1984).

Tothill (1946; 1948) used the presence of shells of aquatic mollusca to demonstrate that late Quaternary floods from the Blue Nile had originally deposited the dark vertisolic clays of semi-arid central Sudan. Williams et al. (1982) analysed the primary shell data obtained by Tothill and plotted the number of mollusc species against depth to show that the permanent water species *Cleopatra bulimoides* was progressively replaced by the semi-aquatic species (with gills and lungs) *Lanistes carinatus*, which was in turn replaced by the terrestrial snail *Limicolaria flammata* some 5,000 years ago (Figure 16.4). Because *Cleopatra* is a mollusc needing permanent water, its distribution across the Gezira and its vertical distribution in the Gezira soil suggest extensive areas of permanent water and seasonal flooding in the eastern Gezira. Both juveniles and adults are present in the clay soil, which led Tothill (1946) to infer temporary flooding each year, followed by the drying out of the water and mass death of all age groups. *Cleopatra* occupied an area at least 180 km long and nearly 50 km wide during the very wet climatic interval between 15 ka and about 9 ka ago (Figure 16.5). As the climate became less wet, the extent of seasonal flooding diminished. Amphibious mollusca, such as *Lanistes* and *Pila wernei*, could survive in the mud and remain moist until the next annual flood. Once the Blue Nile began to cut down into its flood-plain from about 8 ka onwards, the distributary channels responsible for the widespread seasonal flooding would have been beheaded and deprived of any further water. Flooding had ceased by 6–5 ka, although the climate was

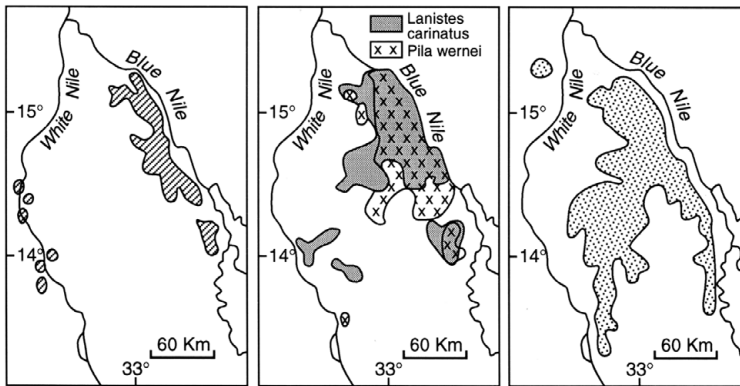


Figure 16.5. Distribution of aquatic (left box) and land (right box) snail shells in Gezira clay, lower Blue and White Nile valleys, central Sudan. (After Williams et al., 1982, fig. 9.7.)

less arid than it is today, allowing the land snail *Limicolaria flammata* to colonise the area. *Limicolaria* is a denizen of the Acacia-Tall Grass savanna, and it is found today south of the 500 mm isohyet in central Sudan. After about 5 ka, the climate became increasingly arid and *Limicolaria* abandoned the Gezira and colonised further south.

More complex analysis involving determining the stable carbon and oxygen isotopic compositions of both aquatic and terrestrial mollusca has enabled more detailed reconstruction of former desert environments. Abell and his colleagues used changes in the $^{12}\text{C}/^{13}\text{C}$ ratio (expressed as $\delta^{13}\text{C}$) and the $^{16}\text{O}/^{18}\text{O}$ ratio (expressed as $\delta^{18}\text{O}$) to determine former temperature and salinity fluctuations in Holocene lakes and springs in the Afar, the Nile Valley and the central Sahara (Abell, 1985; Williams et al., 1987; Abell and Williams, 1989; Abell et al., 1996; Abell and Hoelzmann, 2000; Rodrigues et al., 2000). In addition, Abell and Hoelzmann (2000) and Rodrigues et al. (2000) used changes in the stable oxygen isotopic composition in Nile oyster and Nile gastropod shells to infer changes in rainfall seasonality in north-west Sudan during the early Holocene. In the case of a series of small pans west of the lower White Nile with abundant fossil mollusca in the upper 50 cm of sediment, Ayliffe et al. (1996) were able to show that the source of the precipitation which fell when the pans were full of water some 8,000 years ago was most likely from the South Atlantic and that the pans were fed by local run-off and not by groundwater.

16.5 Plant macrofossils

Plant macrofossils include whole trees, leaves, fruit and charcoal. Care is needed to distinguish between life and death assemblages, because the smaller elements like leaves and fruit can be carried considerable distances by running water, and they may therefore be more indicative of the plants that once grew upstream. Many leaves

and fruit are often beautifully preserved even many millions of years after they died (Hill, 1994a; Hill, 1994b), as the silicified leaves of the late Cenozoic flora in the Upper Lachlan Valley of eastern Australia described by von Ettingshausen (1888) more than a century ago attest. Plant macrofossils and charcoal in archaeological sites provide a partial glimpse into the contemporary prehistoric flora.

A particularly useful form of plant macrofossils are those preserved in packrat or stick-nest rat middens in the drier parts of northern Mexico and the south-west United States (Van Devender and Spaulding, 1979; Betancourt, 1990; Betancourt et al., 1990a; Betancourt et al., 1990b; Cole, 1990; Spaulding, 1990; Van Devender, 1990a; Van Devender, 1990b), as well as central Australia (Pearson and Dodson, 1993; McCarthy et al., 1996; Pearson, 1999). In the deserts of the south-west United States, packrat middens provide a record of environmental change spanning the last 40 ka; in central Australia and the arid Flinders Ranges of South Australia, the stick-nest rat (*Leporillus* spp.) middens investigated so far do not extend back any earlier than the Holocene.

The pollen record preserved in the faecal pellets of *Leporillus* middens from two sites in arid western Australia spanning the last 1,150 years provides information about local vegetation and dietary preferences, in contrast to the regional signal from playa lakes, and suggests a less wooded vegetation cover between 0.9 and 0.3 ka (Pearson and Dodson, 1993). Eight *Leporillus* middens from the arid northern Flinders Ranges indicate wetter conditions and more widespread woodlands between 8.8 and 5.3 ka (McCarthy et al., 1996). The dominance of halophytes (salt-tolerant plants) at the Pleistocene-Holocene transition may indicate continued aridity or a change in rainfall seasonality or more local influences.

The late Quaternary packrat midden record from the arid south-west of the United States and northern Mexico sheds new light on certain critical aspects of desert biogeography and will therefore be discussed here in some detail. The Chihuahuan Desert is an inland continental desert bounded to the west by the Sierra Madre Occidental, to the east by the Sierra Madre Oriental, to the south by the highlands of the Mexican Plateau and to the north by the Rocky Mountains (Van Devender, 1990a). Analysis of 220 packrat middens with 259 associated AMS ^{14}C ages has provided a remarkably detailed picture of vegetation change in this very arid desert during the last 40 ka. Apart from in the lowest parts of the desert, which remained arid, the early Wisconsin climates may have been somewhat wetter than they were in the middle Wisconsin at 31 ka, with more humid conditions during full glacial times at 22 ka. The existence of C_4 perennial grasses indicates rainfall in late spring or summer, when temperatures were relatively warm. The overall LGM climate was mild with few winter freezes, cool summers and higher rainfall throughout the desert. None of the evidence supports the model of a cold, dry LGM climate proposed by Galloway (1970; 1983) and Brakenridge (1978). The demise of the winter rainfall regime took place after 9 to 8 ka, when the modern climatic regime began to be established, and

fully modern climates occurred by 4 ka. Van Devender (1990a, p. 127) also observed that ‘community composition continued to vary subsequently during lesser climatic fluctuations, suggesting that differential responses of plant species to climate changes and continuous variation in climate on several scales have resulted in dynamic plant communities that rarely if ever reach equilibrium (Davis, 1986)’. A corollary to this is that different plants will likely have different lags in responding to any future climatic changes. This short-term Holocene instability is in contrast to the apparently stable early to middle Wisconsin woodland assemblages in certain parts of the Chihuahuan Desert, which persisted without noticeable change for 15,000–20,000 years.

16.6 Plant microfossils

16.6.1 Pollen and spores

Palynology is the study of fossil pollen and spores, and pollen analysis is the study of fossil pollen grains. Pollen analysis was placed on its modern footing by Von Post (1884–1951) in his oral account of Swedish peat bogs in 1916, published two years later (Von Post, 1918), and for many decades it has been the method most widely used to reconstruct terrestrial environments in humid areas during the Quaternary (West, 1977; Faegri and Iversen, 1989; Lowe and Walker, 1997; Williams et al., 1998). It relies on the generally good preservation of fern spores and pollen grains in lake, swamp and bog sediments. Pollen grains have a resistant outer layer composed of a substance called sporopollenin, which enables pollen grains to remain well-preserved in waterlogged deposits. Pollen grains are best preserved in wet, slightly acidic sediments but soon become oxidised and degraded in dry, alkaline environments, which is why they are scarce in desert sediments. Only in the last few decades has pollen analysis been applied with some limited success in arid areas, although problems of low pollen counts and partial preservation remain significant obstacles (Bonnefille, 1972; Bonnefille, 1976; Bonnefille, 1980; Maley, 1980; Maley, 1981; Bonnefille, 1983; Lézine et al., 1990; Bonnefille et al., 2004; Vincens et al., 2007; Lézine et al., 2011). Ritchie et al. (1985) analysed the pollen from a desert lake in northern Sudan and concluded that the savanna vegetation zone extended around 400 km further north during the early Holocene, which is consistent with other evidence from that region (Ritchie and Haynes, 1987).

Pollen grains are produced in the male flowers of gymnosperms (conifers) and angiosperms (flowering plants) and are dispersed by wind, water, insects, birds and other animals. One immediate problem involves distinguishing between near and far sources of pollen. Pollen from aquatic plants is usually deposited quite locally, within the lake, swamp or peat bog in which the plants are growing. However, pollen from plants that rely primarily on wind to disperse their pollen grains can travel many hundreds of kilometres and even thousands of kilometres, as in the case of pollen carried out to sea during dust storms.

Table 16.1. *Processes involved in the production of fossil pollen assemblages from parent plants and subsequent analysis and interpretation. (After Williams et al., 1998, fig. 10.1.)*

-
-
- Pollen production from plants
 - Dispersal
 - Deposition
 - Fossilisation
 - Fossil pollen assemblage
 - Sample collection
 - Pollen extraction
 - Identification and counting
 - Analysis and interpretation
 - Vegetation history and biostratigraphy
 - Environmental (and climatic) reconstruction
-
-

Pollen analysis involves a series of steps (Table 16.1), each of which can add errors to the final interpretation of the vegetation history (Williams et al., 1998, pp. 185–199). Plants first produce pollen, some in great abundance and others less so. The pollen is then dispersed and deposited. Once deposited, some grains will be destroyed by the processes of weathering and erosion, while the remaining grains will become fossilised. These fossil grains comprise the fossil pollen assemblage. The next step is to extract a sample of the pollen-bearing sediment using a variety of coring methods and taking great care to avoid contamination. Then begins the long process of identifying and counting individual pollen grains. The pollen counts are then grouped into pollen zones that are defined according to the relative abundance of different plant species or genera, expressed as a percentage of the total pollen count. Changes in plant assemblages inferred from the pollen are then used to reconstruct the vegetation history. Ideally, the interpretation is calibrated using samples of the modern pollen rain, provided the former plants have living counterparts. The entire process of output-transport-storage-retrieval-sediment preparation-analysis-portrayal-interpretation requires considerable skill and a thorough knowledge of plant ecology. Another possible stumbling block arises when the fossil plant assemblages have no modern counterparts, as Margaret Bryan Davis (1976) so brilliantly demonstrated for the postglacial deciduous forests that colonised North America after the retreat of the great ice sheets some 12,000 years ago.

Pollen analysis enabled each of the European and North American interglacials to be in part defined according to certain diagnostic ferns and other plants. In addition,

it allowed a sequence of postglacial climatic stages to be identified for north-west Europe, including the Older Dryas and the Younger Dryas pollen zones, which were characterised by colder climatic episodes, or stadials, separated by warmer climatic intervals, or interstadials. The *Dryas* is a dwarf shrub native to the arctic and alpine zones of Eurasia and North America and can be seen growing today high in the Swiss Alps and in the mountains of Scandinavia. The alternation of these warmer and colder postglacial climatic phases in high northern latitudes has been confirmed from isotopic analysis of ice cores in Greenland and marine sediments in the North Atlantic. For example, a relatively short cold phase (the Older Dryas) separates two longer interstadials, the Allerød and the Bølling, which are sometimes considered to be a single, complex interstadial, the Bølling-Allerød interstadial (14.6–12.8 ka), which is followed by the cold Younger Dryas stadial (12.8–11.5 ka).

In certain rare cases, the pollen preserved in sites well away from the deserts and desert margins can shed light on past climatic fluctuations within those regions. Lake Suigetsu on the Sea of Japan coast in western Japan offers an excellent example. This lake preserves an annually laminated fine-resolution pollen sequence extending back more than 50,000 years and has been used to calibrate the radiocarbon time scale from 11.2 to 52.8 ka (Bronk Ramsey et al., 2012). Because the lake is located north of the East Asian monsoon front in winter and south of that front in winter, it is very sensitive to changes in Pacific air mass temperature in summer and Siberian air mass temperature in winter. The annual pollen record therefore provides a uniquely detailed history of four separate aspects of climate, namely, winter and summer monsoon intensity, and the respective temperatures of the Pacific air mass in summer and the Siberian air mass in winter (Nakagawa et al., 2006). The Lake Suigetsu winter record shows cooling during the Younger Dryas (YD), which is dated between 12.8 and 11.5 ka in the North Atlantic, but little sign of summer cooling, suggesting that the YD had a greater impact on the Siberian air mass than it did on the Pacific air mass. From this Nakagawa et al. (2006) concluded that the monsoon front provided a major paleoclimatic boundary that divided the Northern Hemisphere into several distinct blocks, each showing different responses to changes in ocean circulation, with the YD missing or attenuated south of the monsoon front. Nakagawa et al. (2003) had previously observed that climatic changes in the North Atlantic and Japan were not synchronous, with warming starting earlier in Japan (at 15 ka) and later in the North Atlantic (at 14.6 ka, the onset of the Bølling-Allerød interstadial) and cooling in Japan (12.3–11.25 ka) lagging the YD cooling in the North Atlantic (12.8–11.5 ka) by 250 to 400 years.

Pollen is not only preserved in Quaternary sediments but has also been used to study much older environments, including those dating back to the Early and Middle Cenozoic. For example, Alley and Beecroft (1993) analysed fossil spore and pollen assemblages, as well as foraminifera, in the Eucla Basin of South Australia to

reconstruct paleochannel activity and Eocene sea level fluctuations in this presently semi-arid area. Alley (1998) and Alley et al. (1996; 1999) extended this investigation further north to include the Lake Eyre Basin and carried out detailed evaluations of the nature and timing of major phases of deep weathering, ferricrete formation, silicification, paleochannel aggradation and vegetation changes. They concluded that deep weathering was prevalent before channel sedimentation and was possibly as old as the early Mesozoic. They were able to use pollen evidence to reconstruct the types of forests growing in central and southern Australia during the Cenozoic. Temperate rainforest grew along the southern continental margin during the Palaeocene. The rainforest was replaced by open woodland during the Oligocene-Miocene. At this time, shallow alkaline lakes occupied parts of the paleochannels in the inland reaches, supporting a diverse fauna, including crocodiles. By Pliocene times, some 5 million years ago, continued desiccation gave rise to an environment characterised by chenopod shrubland and open woodland.

Helene Martin (2006) put together all the pollen evidence obtained in the last fifty years from boreholes throughout Australia by herself and others and was able to provide a comprehensive overview of the development of arid vegetation in Australia and the associated changes in Cenozoic climates. Her conclusions were entirely consistent with the evidence from phylogenetics and phylogeography analysed by Byrne et al. (2008), which confirmed that the onset of aridity in Australia is first evident in the mid-Miocene plant record some 15 million years ago.

In the drier mountainous regions of southern Africa, the pollen preserved in rock hyrax dung in rock shelters has been dated by ^{14}C and has provided a partial but still useful record of late Quaternary vegetation changes that was not obtainable by other means (Scott and Woodborne, 2007). The ^{13}C content of the dung has provided additional paleoclimatic information (Scott and Vogel, 2000). Longer records of vegetation change have been obtained from the relatively sparse pollen preserved in lake and pan sites in South Africa (Scott et al., 1995; Scott, 2002; Scott et al., 2003; Scott et al., 2008), including a 190 ka record for the Savanna Biome from Tswaing Crater near Pretoria (Scott, 1999).

The late Quaternary record of vegetation changes in semi-arid north-west Australia comes not from deposits located on land but from marine sediment core Fr10/95-GC17 collected about 60 km west of the Cape Range Peninsula ($22^{\circ}2.74'\text{S}$, $113^{\circ}30.11'\text{E}$, water depth 1,093 m) (van der Kaars and De Deckker, 2002). The core site is close to the present-day southern margin of the Australian summer monsoon and at present receives 200–300 mm of annual rainfall. Pollen data from this core provides a sensitive record of changes in the latitudinal position of the monsoon. From 35 to 20.4 ka, herbs or small shrubs dominated the regional vegetation and there were relatively few trees, pointing to a significant reduction in summer rainfall. This was the driest period of the last 100 ka, which confirms previous interpretations (De Deckker et al., 2002) that the summer monsoonal regime and associated precipitation

failed to reach the north-west of the continent during this time. Wetter conditions resumed at 14.2 ka both here and elsewhere in the seasonally wet tropics of northern Australia.

16.6.2 Diatoms

Diatoms are tiny, single-celled phytoplankton, or, more technically, eukaryotic microalgae with a siliceous skeleton. They live in lakes, marshes, rivers and the sea. They are very good indicators of water chemistry, depth and temperature because many individual species are sensitive to even slight changes in these factors. The uptake of dissolved silica by diatoms ensures that the siliceous diatom cells, or frustules, accumulate on the lake floor as resistant, usually well-preserved fossils. The frustules of each diatom species vary in shape, size and ornamentation, very much like pollen grains. Diatom species are grouped into assemblages and zones, and they can be used to interpret past environmental changes on the basis of known species tolerances to water salinity, temperature and depth (Gasse, 1975; Gasse, 1980; Smol et al., 1986; Gell, 1997; Chalié and Gasse, 2002). Another approach still being developed is the analysis of the stable oxygen isotopes contained within the silica (SiO_2) of the frustules (Leng and Barker, 2006; Leng and Barker, 2007; Leng and Sloane, 2008).

Gasse (1975) carried out a pioneering study of the diatom assemblages of Pleistocene and Holocene lakes in the hyper-arid Afar Desert of Ethiopia and was able to distinguish those lakes fed mainly by surface run-off from those fed only from groundwater inflow. She found that fluctuations in the first type of lake were a result of fluctuations in rainfall and evaporation in the upland headwaters, whereas the lakes fed from groundwater were comparatively insensitive to regional climatic fluctuations.

In the Chad Basin, the onset of late Pliocene aridity (Servant, 1973; Sepulchre et al., 2006) was associated with a diatom flora in the lake deposits that was indicative of cooler-than-present temperatures (Servant and Servant-Vildary, 1980). The diatom and pollen content of Pliocene Lake Gadeb in the south-east uplands of Ethiopia (see Chapter 11) also indicates cooling and progressively drier conditions around 2.5 Ma ago (Williams et al., 1979; Gasse, 1980; Bonnefille, 1983), when the Northern Hemisphere ice caps began to expand. This was a time of widespread intertropical cooling and desiccation that was most likely linked to the expansion of the Northern Hemisphere ice caps.

Diatoms can often reveal a remarkably detailed picture of past climatic fluctuations in deserts. The diatom assemblages and stable isotopic composition of Holocene lake sediments in the northern Sahara show very rapid changes in water chemistry from fresh to highly saline within this time interval (Fontes et al., 1985). Analysis of the pollen, diatoms, ostracods and stable isotopes preserved within the sediments of five lakes in the arid Qinghai-Tibetan Plateau region and northern Xinjiang in north-west

China indicates an abrupt increase in summer rainfall at 12.5–11 ka that lasted until 8–7 ka, with maximum aridity at all five lake sites from 4.5 to 3.5 ka (Fan et al., 1996; Gasse et al., 1996, Van Campo et al., 1996; Wei and Gasse, 1999).

16.6.3 Other plant and animal microfossils: phytoliths, charophytes and sponge spicules

A number of other plant and animal microfossils can provide supplementary information about past desert environments. *Phytoliths* are the microscopic remains of the particular parts of certain plants that are composed of silica, and they can be used to identify certain terrestrial plants. *Charophytes* are green algae. The reproductive part of the organism is called the oogonium, and calcareous charophyte oogonia have been used to date late Pleistocene lake lunettes in semi-arid New South Wales (Williams et al., 1991a).

Williams et al. (1986) and Adamson and Williams (1987) used the presence or absence of diatoms, phytoliths and sponge spicules in Pliocene fluvio-lacustrine deposits in the Middle Awash Valley of the Afar Desert to distinguish between fluvial and lacustrine beds. Barboni et al. (1999) analysed the phytoliths within these Pliocene deposits in far greater detail, as an adjunct to the pollen analysis by Bonnefille et al. (2004), which showed a mosaic vegetation pattern of woodland, grassland and dense riparian forest during the time that the Pliocene hominid *Australopithecus afarensis* roamed this area.

Sponge spicules are microscopic cylinders composed of silica. The discovery of sponge spicules that were used to temper pottery in the White Nile Valley 2,000 years ago came from a site situated around 350 km north of the nearest present-day swamps (Adamson et al., 1987a), and confirmed the account of these swamps given to the Emperor Nero after his centurions found their attempt to trace the source of the White Nile blocked by impenetrable swamps (Mawson and Williams, 1984).

16.7 Conclusion

The evidence from fossil plants and animals has been an invaluable supplement to other lines of evidence relating to climatic change in deserts, and in many instances it is the only evidence available. The fossils of large and small vertebrates afford reliable insights into past habitats, but care is needed in distinguishing between remains that have been transported by running water, humans or other predators and remains that are indeed in primary context. Invertebrate fossils such as freshwater mollusca, ostracods, cladocera and chironomids are valuable indicators of water depth, temperature and salinity or alkalinity, and if they are used together, they can form a powerful tool to reconstruct past changes in desert lakes and swamps. Among plant fossils, pollen and spores have been most widely used to reconstruct past changes in vegetation

throughout the Cenozoic. Macroscopic plant remains preserved in packrat middens have provided a uniquely detailed picture of late Quaternary vegetation changes in the deserts of northern Mexico and the south-west United States. Out of this work has come the realisation that the responses of the biota to climatic change can be quite variable, with some species showing a synchronous response, others a delayed but sudden (or step-function) response and others showing a delayed and time-transgressive (or diachronous) response. Paradoxically, the fossils that provide the best overall chronologic control of global climatic changes are not found in the deserts but in the oceans, where the marine foraminifera have provided a detailed record of Cenozoic changes in global temperature and of the Quaternary glacial-interglacial cycles which had such a profound effect on desert environments.

Prehistoric occupation of deserts

Ancient life in deserts was in some ways distinctively different from that observed in historic adaptation to arid lands, reinforcing at a global level the inference that historically observed desert societies and adaptations are relatively recent consequences of evolving systems.

Peter Veth, Mike Smith & Peter Hiscock

Desert Peoples: Archaeological Perspectives (2005, pp. 79–80)

Entrer dans l'ère néolithique, c'est pénétrer dans un monde en mutation capitale, sans jamais voir disparaître un certain savoir-faire parfois rattaché à un certain savoir-vivre.

To enter the Neolithic era is to penetrate a world undergoing major change while retaining a certain element of acquired know-how linked on occasion to an enhanced appreciation of ways of living.

Lionel Balout & Colette Roubet

The Sahara and the Nile (1980, p. 169)

17.1 Introduction

In [Chapter 16](#) we considered how plant and animal fossils have been used to reconstruct past changes in climate in deserts and desert margins. In this chapter, we focus on a very particular set of fossils – those of our prehistoric ancestors ([Figure 17.1](#)). This record is unique for two main reasons: we cannot only analyse the fossil evidence to see what it can tell us about past human behaviour such as gait and diet, but we can also examine the stone tools and other remains of prehistoric human activity to bolster the often sparse evidence that can be derived from the fossil bones. In addition, we can use the insights gained from recent advances in molecular biology to help reconstruct the pattern and timing of prehistoric human migrations (Cooper and Stringer, [2013](#)). We therefore need to study both stone tools and hominid fossils, as well as any other associated material. Following Brunet et al. ([2005](#), p. 753), the term *hominid* is used here to refer to ‘all taxa that are closer to humans than chimpanzees’

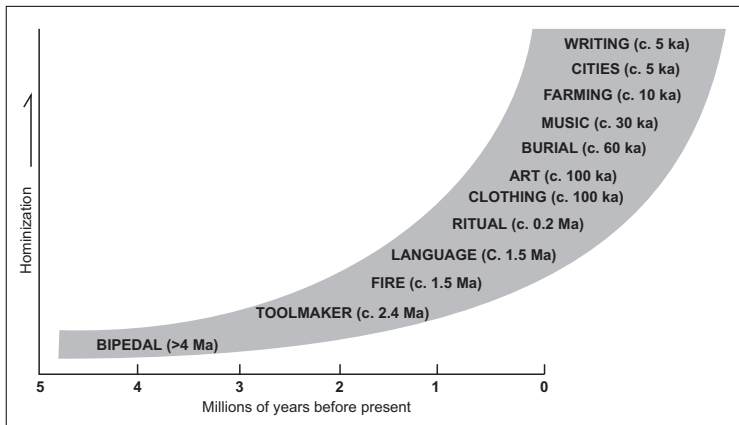


Figure 17.1. The development of human culture during the late Pliocene and Quaternary, showing increasing ‘hominization’ through time. (After Williams et al., 1998, fig. 11.6.)

and so covers humans and their discernible ancestors, as inferred from changes in skeletal morphology. The current use of *hominin* as a substitute for hominid remains a matter of taste, so we opt for the already well-known term hominid, a preference endorsed also by White (2010, p. ii).

The aims of this chapter are to trace the history of early human occupation of the presently arid and semi-arid regions of the world and to examine the interactions between prehistoric peoples and their environments. We then assess what light archaeological evidence has shed on former environmental and climatic changes in deserts. Finally, we discuss the potential use of prehistoric stone tool assemblages as ‘zonal fossils’ in order to establish a relative chronology of past environmental fluctuations when other means are lacking.

17.2 Late Miocene and Pliocene hominid fossils of Africa

Charles Darwin was probably the first naturalist to realise that Africa was the continent most likely to yield fossil remains of early humans. In his book *The Descent of Man*, he argued that:

In each great region of the world the living mammals are closely related to the extinct species of the same region. It is therefore probable that Africa was formerly inhabited by extinct apes closely allied to the gorilla and chimpanzee; and as these two species are now man’s nearest allies, it is somewhat more probable that our early progenitors lived on the African continent than elsewhere. (Darwin, 1871, p. 570)

He went on to point out that ‘the discovery of fossil remains has been a very slow and fortuitous process. Nor should it be forgotten that those regions which are most likely

to afford remains connecting man with some extinct ape-like creature, have not as yet been searched by geologists' (op. cit., p. 571).

Darwin's inference as to the close relationship between early hominids and early African primates has been confirmed by modern genetic studies which indicate that divergence between ancestral chimpanzees and ancestral hominids may have occurred between 7 and 5 Ma ago (Pilbeam, 1986), and possibly even earlier. Caution is advised before accepting such ages as immutably reliable, because we cannot assume that mutation rates have been constant over time. We therefore need to bear in mind that the 'molecular clock' pioneered by Sarich and Wilson (1967) will always require calibration against independently dated elements of the fossil record using the techniques outlined in Chapter 6. The stunning continuing discoveries of very late Miocene, Pliocene and younger hominid fossils in limestone caves in semi-arid southern Africa (Dart, 1925; Partridge et al., 2003; Walker et al., 2006), at Olduvai Gorge in semi-arid Tanzania (Blumenschine et al., 2003), around Lake Turkana in arid northern Kenya (Spoor et al., 2007), from the Chad Basin (Brunet et al., 1995; Brunet et al., 2002; Brunet et al., 2005) and from the Middle Awash Valley in the otherwise hyper-arid Afar Desert (White et al., 1994; Alemseged et al., 2006; White et al., 2006; Wynn et al., 2006) have fully vindicated Darwin's cautious prediction. Indeed, Pliocene hominids appear to be unique to Africa, with later migrations from Africa to Eurasia not occurring until well after the development in Africa of the first stone tools some 2.5 Ma ago (Corvinus, 1975; Roche, 1980; Semaw et al., 1997; Roche et al., 1999; Ron and Levi, 2001; Balter, 2002).

The question has often been asked as to why Africa was such a good place for Pliocene hominids to inhabit (see, for example, Bishop, 1978). We should perhaps widen the question and ask why certain parts of Africa were also good places in which to die, to be preserved and to be found again millions of years later (Coppens et al., 1976; Hay, 1976; Rapp and Vondra, 1981). In the case of the Afar Desert and other arid sectors of the East African Rift Valley, it can be argued that bone preservation was aided by rapid burial in a dry, alkaline environment, with soft sediments protected beneath younger basalt flows. Later exposure and discovery were made possible in these tectonically active regions by uplift and faulting, which have been the object of intensive study over the past forty years (Taïeb, 1974; Pilger and Rösler, 1976; Accademia Nazionale dei Lincei, 1980; Rapp and Vondra, 1981; Popoff and Tiercelin, 1983; Adamson and Williams, 1987; Beyene and Abdelsalam, 2005; Chorowicz, 2005; Yirgu et al., 2006; Corti, 2009).

The earliest hominid fossils presently known come from the Djurab Desert in the northern Chad Basin near the Toros-Menalla fossil-rich site, and they are associated with a savanna fauna (Brunet et al., 1995; Brunet et al., 2002; Brunet et al., 2005). Here, Michel Brunet and his colleagues have recovered the remains of a very late Miocene hominid they termed *Sahelanthropus tchadensis*, with a cranial capacity of about 360 cc. They considered the hominid to be around 7 Ma in age from the associated

fauna, confirmed at 7.2–6.8 Ma on the basis of twenty-eight $^{10}\text{Be}/^9\text{Be}$ cosmogenic nuclide ages (Lebatard et al., 2008). This hominid also shows possible evidence of an upright posture (Zollikofer et al., 2005). In arid northern Kenya, Pickford and Senut (2001) discovered the remains of another hominid around 6 Ma old which they named *Orrorin tugenensis* (Senut et al., 2001). The femur of *O. tugenensis* shows that it was bipedal but no more closely related to *Homo* than to *Australopithecus* (Richmond and Jungers, 2008). However, by far the best-dated hominid fossils are the Australopithecines from the Afar Rift, from Olduvai Gorge and from certain limestone caves in South Africa, which range in age from around 4.2 Ma to perhaps 1.0 Ma (White et al., 2000; White et al., 2006; Gibbons, 2011; Green and Alemseged, 2012; Haile-Selassie et al., 2012; Lieberman, 2012).

Australopithecus afarensis was certainly capable of walking upright, as the 3.6 Ma footprints excavated by Mary Leakey and preserved in carbonatite ash at Laetoli in Tanzania so eloquently attest (Leakey and Hay, 1979). The Bahr el Ghazal (Arabic for ‘river of the gazelles’) Valley in the eastern Chad Basin has yielded fossil remains of *Australopithecus bahrelghazali* (Brunet et al., 1996), and the associated fauna suggest an age of around 3.5–3.0 Ma, which is confirmed by a single ^{10}Be age of 3.58 ± 0.27 Ma (Lebatard et al., 2008), indicating that these remains were contemporary with *A. afarensis* in East Africa.

A group of well-preserved hominid fossils placed within the new genus *Ardipithecus* have been recovered from the Afar Desert. They are somewhat older than the Australopithecines, being bracketed between 5.8 and 4.4 Ma, and are considered by their finders to be the oldest unequivocal hominids from the Middle Awash Valley in the Afar Rift of Ethiopia (White et al., 1994; Renne et al., 1999; Haile-Selassie, 2001; WoldeGabriel et al., 2001), a view increasingly accepted by workers in this highly competitive field of natural science. The Australopithecines comprise ‘all Pliocene hominid taxa that do not belong in the genera *Ardipithecus* and *Homo*’ (Brunet et al., 2005, p. 753). They had small brains relative to body mass and teeth that were often well-adapted to a dominantly vegetarian diet (e.g., *Australopithecus robustus*/*Paranthropus robustus*) in South Africa and *A. boisei*/*Paranthropus boisei* in East Africa). It is not known if the Australopithecines made stone tools, although there is some sparse initial evidence that *Australopithecus afarensis* may have used them to cut meat from several bones at the Dikika site near Gona and Hadar in the western Afar Desert 3.4 Ma ago (McPherron et al., 2010).

With the emergence of stone tool-making some 2.5 Ma ago, we see a progressive increase in both cranial capacity and brain size relative to body mass (Figure 17.2). Louis Leakey suggested that *Homo habilis* (around 2.3 to 1.4 Ma) was the first stone tool-maker of any consequence, followed by *H. erectus* after about 1.4 Ma (Leakey, 1966). However, recent work at Ileret, east of Lake Turkana in arid northern Kenya has shown that *H. habilis* and *H. erectus* coexisted in the same lake basin for nearly half a million years (Spoor et al., 2007). Both in the Middle Awash region of the Afar

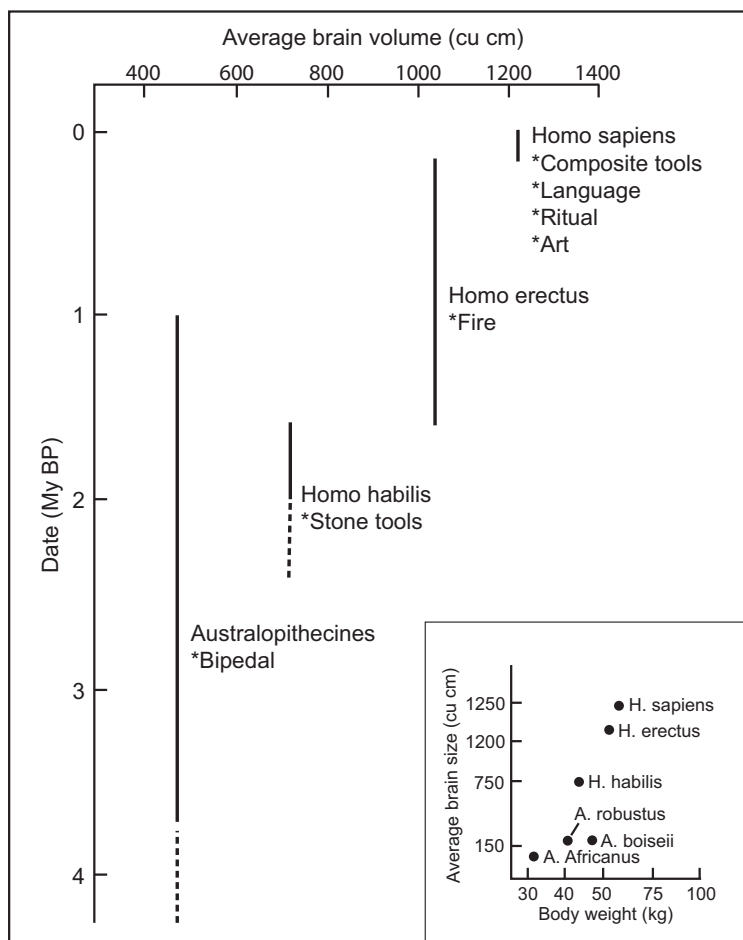


Figure 17.2. Changes in hominid physical and cultural development from early Pliocene to late Pleistocene. (After Williams et al., 1998, fig. 11.3.) The inset shows that the *Homo* line has a much higher ratio of brain size to body weight than any of the Australopithecines had.

Rift in Ethiopia and in now arid northern Kenya, remains of *H. erectus* have been found in association with abundant Acheulian stone tools (Asfaw et al., 2002). Later work saw several new species placed on the hominid family tree (family ‘bush’ might well be a more appropriate metaphor), including *H. rudolfensis* from the arid Lake Turkana region in northern Kenya and *H. ergaster*, also from northern Kenya, with an age range between around 1.9/1.8 and 1.5/1.3 Ma.

Hominid taxonomy varies according to whether the proponents are lumpers or splitters, and we would do well to remember that more than 5 million years of geological time are represented by only a few hundred often highly incomplete fossil remains. Wood and Collard (1999), Gibbons (2002) and Wood (2006) provide useful

summaries of the complex field of hominid taxonomy, and no doubt amended versions will follow. However, we can be reasonably confident that the brains of *H. erectus* and its variably named successors (*H. antecessor*, *H. rhodesiensis* (0.7–0.3 Ma), *H. heidelbergensis* (0.5–0.3 Ma) and *H. neanderthalensis* (around 300 to 35 ka)) became progressively larger and more complex, attaining a cranial capacity of about 1,400 cc by around 0.3 Ma ago (Stringer and McKie, 1996). The as yet poorly dated Denisovans may have arisen before 300 ka ago and appear to have interbred with both Neanderthals and *H. sapiens* (Cooper and Stringer, 2013). Anatomically, modern humans (*H. sapiens*) are known from the lower Omo Valley in southern Ethiopia in sediments dated between 195 and 104 ka, with the earlier date more likely (McDougall et al., 2005; Fleagle et al., 2008; McDougall et al., 2008), and from South Africa in cave deposits at Border Cave and Klasies River Mouth dated to at least 100 ka (Grün and Stringer, 1991; Stringer and McKie, 1996; Bräuer et al., 1997).

The hominid discoveries are of interest in their own right, and learning about them appeals to our atavistic desire to understand more about our own past. However, in regard to our main theme – climatic change in deserts – the hominid fossils alone are of secondary interest. What is of interest for our present purpose is the huge volume of cognate research devoted to unravelling the environmental context of the fossil discoveries (Taïeb, 1974; Hay, 1976; Pickford, 1994; WoldeGabriel et al., 1994; Kalb, 1995; Barboni et al., 1999; WoldeGabriel et al., 2001; Quade et al., 2004; Wynn et al., 2006). For example, pollen analysis of the late Pliocene fossil-bearing deposits in the Middle Awash Valley has revealed a mosaic vegetation pattern of woodland, grassland and dense riparian forest (Bonnefille et al., 2004), which has been confirmed by analysis of the stable oxygen and carbon isotopic composition of pedogenic carbonate nodules within the fossil soils associated with the hominid fossil discoveries (WoldeGabriel et al., 2009).

One tantalising question relating to the origin of the African hominids concerns the extent to which their evolution was influenced or even determined by climatic changes, especially late Cenozoic cooling and desiccation which led to an expansion of savanna grassland at the expense of forest and woodland (Brain, 1981a; Brain, 1987; Vrba et al., 1995; deMenocal, 2004; Derricourt, 2005; Behrensmeyer, 2006; Trauth et al., 2007; Trauth et al., 2010; Cerling et al., 2011; deMenocal, 2011; Maslin and Christensen, 2007). Changes in African faunal assemblages, especially bovids, indicative of more open habitats at 2.9–2.6 Ma and between 1.9 and 1.6 Ma appear to coincide with step-function changes in hominid evolution and the emergence of the genus *Homo* (deMenocal, 2004; Maslin and Christensen, 2007; deMenocal, 2011). Maslin and Christensen (2007) and Trauth et al. (2007; 2010) emphasised the extreme climatic variability at 2.7–2.5, 1.9–1.7 and 1.1–0.7 Ma, and they suggested a causal link between these interludes of high environmental variability and speciation and dispersal episodes among hominids and other mammals in East Africa.

17.3 Stone Age cultures and prehistoric stone tools

Klein (1989) has pointed out that stone tools are defined primarily on the basis of their shape and not their actual function, about which we can often only surmise, and so it is very hard to infer former human behaviour purely on the basis of stone tool morphology. To do this more convincingly, we need to draw on the evidence left by tool-use wear and by plant and animal residues on the cutting edges of stone tools, supplemented by tool-making and tool-use experiments designed to test any propositions arising from the purely archaeological remains (Keeley, 1980; Keeley and Toth, 1981; Schick and Toth, 1995). Ethnographic evidence derived from modern hunter-gatherer societies provides another fruitful means of formulating testable hypotheses or models of past human behaviour (Clark, 1980). Needless to say, not all tools are made of stone. Observations of wild chimpanzee behaviour suggest that the early hominid use of sticks for digging and of sticks and other plant materials for making nests or shelters was likely (Goodall, 1976). The use of fire-hardened wooden spears may extend well back in the prehistoric record, although suitable conditions for their preservation are rare (Clark, 1969), and hafting stone points to wooden shafts to make spears has a respectable antiquity of around 500,000 years in southern Africa (Wilkins et al., 2012), which is 200,000 years more than had been previously thought (e.g., Ambrose, 2001).

The earliest stone tools are pebbles from which several flakes have been struck. These pebble tools (Figure 17.3) are found across Africa, and the oldest ones presently known (2.6–2.5 Ma) come from the Gona Valley in the Afar Desert of Ethiopia on the left bank of the Awash River, where they have been dated using a combination of geomagnetic polarity stratigraphy and argon/argon dating of vitric tephra beds (Semaw et al., 1997). These tools show evidence of several generations of flake scars, and the ‘large number of well struck flakes with conspicuous bulbs of percussion’ indicates that the toolmakers had ‘a clear understanding of conchoidal fracture mechanics’ (Semaw et al., 1997, p. 335). At Hadar on the left bank of the Awash River, stone tools and *Homo* fossils occur together and have an age of 2.33 ± 0.07 Ma (Kimbél et al., 1996). The name given to this pebble tool tradition is *Oldowan* from their early discovery at Olduvai Gorge in Tanzania by Mary and Louis Leakey, where they were dated to around 1.8 Ma (Leakey et al., 1961; Evernden and Curtis, 1965).

About 1 million years after the inception of stone tool-making at around 2.5 Ma, and apparently quite abruptly at 1.6–1.5 Ma ago, our hominid ancestors contrived to detach large flakes about 15–20 cm long from big blocks of rock (Ambrose, 2001). These flakes were then struck with a stone hammer, either on both sides or on a single side, to form a bifacially or unifacially worked large stone flake, termed a biface or a uniface. A biface with a point at one end is often called a hand-axe; those with a chisel-like edge are called cleavers. Together with flakes and spheroids, they characterise what is known as the *Acheulian* cultural tradition (Figure 17.3), from the

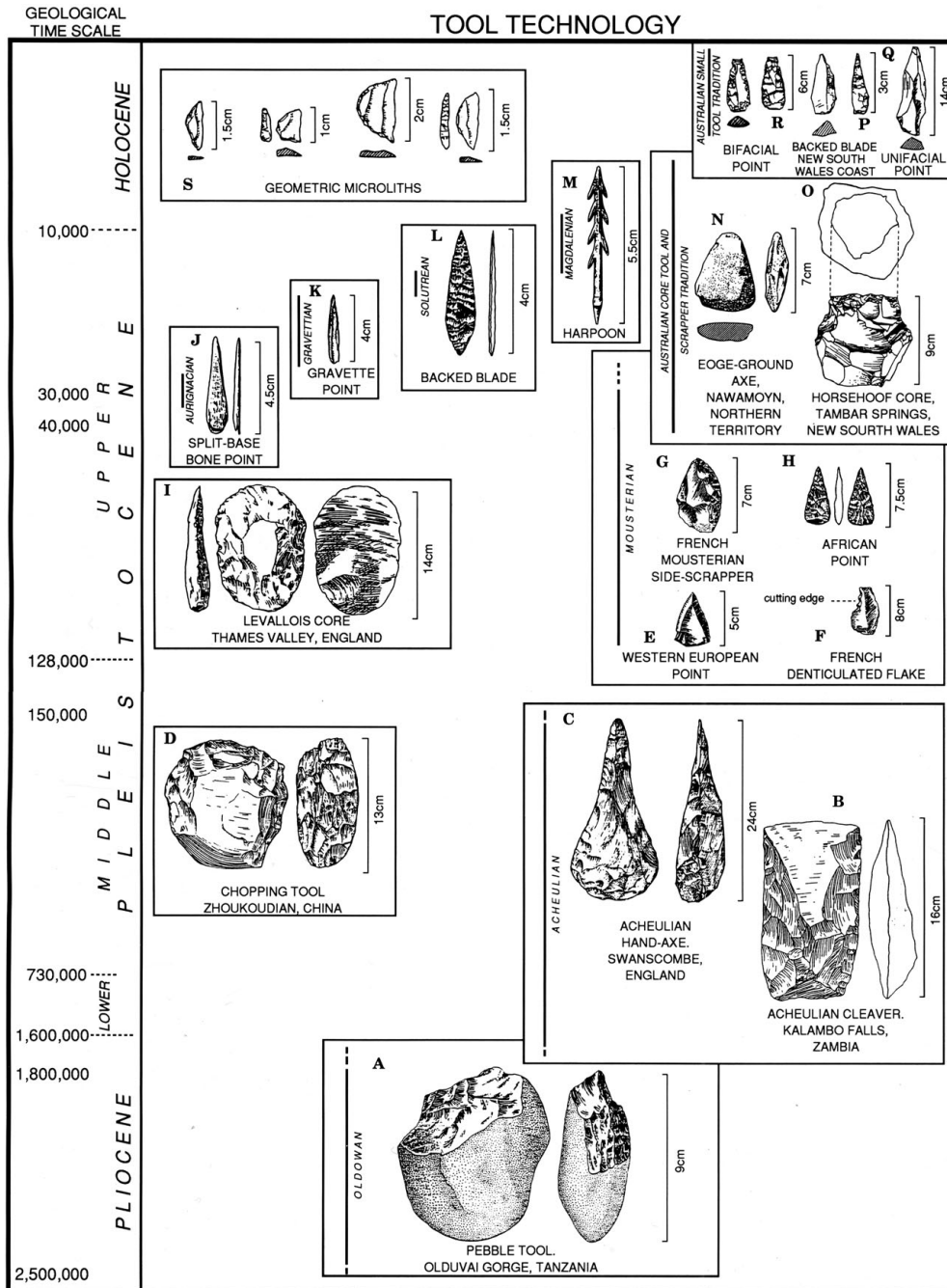


Figure 17.3. Origin and development of Palaeolithic stone tool technology from 2.5 Ma to 10 ka. (After Williams et al., 1998, fig. 11.4.) Neolithic polished stone tools are not shown here but are shown in Figure 17.5.

river valley near Saint Acheul in northern France where the French customs officer and amateur archaeologist Jacques Boucher de Perthes first excavated them in the early nineteenth century. The Oldowan and Acheulian traditions together make up the *Lower Palaeolithic* or *Early Stone Age* (ESA). For simplicity, the terms Lower Palaeolithic and Early Stone Age are used synonymously, although the former applies strictly to Eurasia and Africa north of the Sahara and the latter only to Africa south of the Sahara, because this desert was earlier thought to be an effective barrier to human movement, which was certainly not the case during the wet phases of around 120–110 and 50–45 ka (Osborne et al., 2008; Castañeda et al., 2009).

The transition from Early to *Middle Stone Age* (MSA) is as yet poorly dated and may have taken many thousands of years. Previous best estimates placed the ESA/MSA transition in Africa at around 300 ka (Ambrose, 2001), although as noted earlier in this section, an age of around 500 ka now seems possible (Wilkins et al., 2012). In Asia and Europe, the transition from Lower to *Middle Palaeolithic* is probably of similar antiquity (Figure 17.3). In Qesem Cave in Israel, uranium-series ages suggest a long transition between the Lower Palaeolithic Acheulian and the Middle Palaeolithic *Mousterian*, starting before 382 ka and ending around 200 ka ago (Barkai et al., 2003).

A hallmark of the MSA is what is termed the *Levallois* technique, which involved fashioning a stone *core* by striking off flakes parallel to the long axis of the core (Ambrose, 2001, fig. 3). (A core is what remains of the parent stone once it has been flaked). The resulting relatively thin flakes were then reworked to form blades, some of which were pointed and attached to wooden shafts to form spears (Figure 17.3). These Mousterian points are diagnostic of a hunting tradition. Hafting and core preparation were two technological innovations of the Middle Palaeolithic/MSA. Hafting and the use of spears with sharp stone points allowed for more efficient forms of hunting, while initial core preparation enabled production of a greater number of blades with cutting edges per unit volume of stone. There is evidence of regional specialisation during the MSA and of a more diverse array of stone tools, which allowed for more effective use of a greater variety of natural resources (Clark, 1980; Clark, 1982; Klein, 1989; Van Peer, 1998). It was during the late Acheulian and the MSA that habitats that had been previously avoided began to be occupied, at least intermittently, including tropical deserts and rainforests.

Around 50–40 ka, we see another major cultural change associated with the use of a greater range of tool-making materials, such as bone, antler, shell and ivory. The advent of the *Upper Palaeolithic*, or *Later Stone Age* heralds the proliferation in certain regions of rock art (paintings, as well as engravings), sculpture and the fashioning of stone artefacts of great beauty for aesthetic purposes rather than for purely utilitarian daily use (Figure 17.3). In the drier parts of Africa, perforated ostrich eggshell beads become common. High quality sources of stone (such as certain types of obsidian in the Kenya Rift Valley) were sought from much further away than they had been previously. In India, Africa and Australia, certain types of siliceous rock were heated

first in order to make them easier to flake and work. Trade over great distances (hundreds of kilometres) became more common and the exchange of goods allowed social networks and reciprocal alliances to be developed, providing essential insurance for desert dwellers in times of drought. The LSA tool-kits are highly diverse and often very specialised.

Two items in the Upper Palaeolithic/LSA tool-kit were later to have important repercussions for humanity as a whole, namely, sickles and grindstones. The sickles were used to harvest wild cereal grasses, and consisted of a handle (of bone, wood or antler) and small, sharp worked flakes, or *microliths*, attached to the sickle handle with gum, resin or other adhesive. The grindstones generally consisted of a lower grindstone shaped from a wide, upwardly concave slab of sandstone or other suitable rock and a smaller upper grinder that was often roughly cylindrical in cross-section. An alternative to stone grinders, and one still widely used in the drier regions of Africa and Asia, consists of a deep, hollowed-out wooden mortar fashioned from a tree branch or trunk and a solid length of polished wood about 10 cm in diameter and up to 1.5 m long used as a pestle to grind roots and seeds. A third innovation that appeared towards the end of this period, in what is variously called the *Mesolithic*, or *Epi-Palaeolithic*, was the discovery of firing clay to make earthenware pots that could be used to store food items safely away from rodents or to cook food. The significance of these three items lies in the roles they played in pre-adapting human societies to the eventual harvesting and storage of domesticated cereal grains – the signature of the Neolithic.

The *Neolithic*, or ‘new stone age’, began about 11,000 years ago and is characterised by the independent domestication of plants and animals at a few key localities in the Near East (the so-called ‘Fertile Crescent’), China, India-Pakistan, West Africa, Mexico and South America. Once human societies began to domesticate and herd suitable animals, such as goats, sheep and cattle (Figure 17.4), the practice spread very rapidly and, within a few thousand years, cultivating crops and herding domestic animals became the norm in most parts of the world, including the deserts and desert margins (Clark and Brandt, 1984; Diamond, 1998). To go from collecting and storing wild cereal grains to harvesting and storing their domesticated equivalents is relatively straightforward, as Ann Stemler pointed out more than thirty years ago (Stemler, 1980). The prerequisites are, first, an efficient harvesting tool, which was already in use at the end of the Upper Palaeolithic/LSA in the form of the sickle. Second, early farmers needed to collect, store and later sow the heads of mutant wild plants in which the abscission zone at the base of the inflorescence fails to function and leaves the cereal grains or wild grass seeds on the plant until late in the growing season instead of releasing them intermittently, as befits the reproductive survival strategies of wild grasses.

The herds of domestic animals provided the Neolithic inhabitants of the arid and semi-arid lands of Eurasia, Africa and South and Central America with milk and

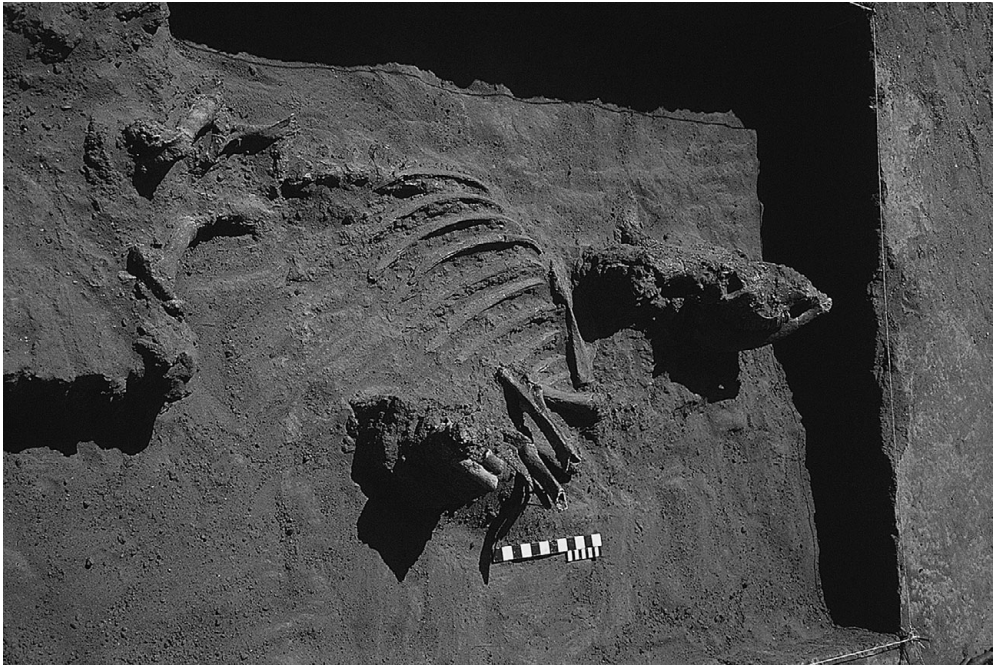


Figure 17.4. Short-horned Neolithic cow skeleton (*Bos brachyceros*), Adrar Bous, south-central Sahara.

meat for sustenance and wool and hides for shelter and clothing. A diet of cereal porridge and milk is easier for infants to digest than their probable pre-Neolithic diet, so infant mortality rates decreased and birth rates increased. The Y chromosome genetic evidence shows an expansion of East Asian Mongoloid groups around 7 ka, perhaps as a result of millet and rice farming, a situation matched in the Near East after farming arose there (Underhill et al., 2001). More than thirty years ago, May (1978) estimated that the world population at the start of the Neolithic some 10,000 years ago amounted to little more than about 5 million, increasing to 100 million by 5 ka, and thereafter increasing ever faster to around 500 million by 300 years ago, 1 billion (1×10^9) by 1850, 4 billion by 1978, and more than 7 billion by 2012. With more people came an ever greater and often adverse impact on ecosystems, including soils and water (Diamond, 2005), as well as increasing pollution of land, air and water, topics we discuss in Chapters 24 to 26.

Another feature of Neolithic life was also to have lasting repercussions. With the ability to obtain and store a substantial food surplus, there was a progressive change from nomadic herding to sedentary farming, with the growth of villages and, ultimately, major urban centres in well-watered river valleys or upland sites with reliable supplies of water. Trade proliferated, as did the emergence of social hierarchies, with a small, powerful and wealthy ruling class, supported by artisans, priests and soldiers.

Writing on clay tablets or papyrus sheets probably developed from the Neolithic use of trade-marks, and various forms of alphabets and calendars arose among the later metal-working cultures of Eurasia and the Americas, none of which were associated with desert hunter-gatherer societies, among whom there was no perceived need for such oddities.

17.4 Prehistoric occupation of the deserts and semi-deserts

Analysis of mitochondrial DNA (mtDNA) in extant human populations confirms movement out of Africa on a number of occasions during the past 1.5 million years or more (Cann et al., 1987; Underhill et al., 2000; Cann, 2001; Hammer et al., 2001). Initial movement out of Africa is associated with small bands of *H. erectus*, who were equipped with a relatively unspecialised Acheulian tool-kit and the ability to make and use fire (Clark, 1975; Clark and Harris, 1985; Goren-Inbar et al., 2004). Earlier *Homo erectus*/*Homo ergaster* groups may have brought their Oldowan technology to Eurasia (Carbonell et al., 1999), as evident at Dmanisi in Georgia (Vekua et al., 2002), where a complete *Homo* skull has an age of around 1.8 Ma (Lordkipanidze et al., 2013). This suggests that these early migrants were able to obtain enough in the way of plant and animal foods for survival but that their overall impact on their habitat was minimal. One probable migration route was across the Sinai and southern Negev and then across the Dead Sea Rift into Arabia and thence to Asia (Ron and Levi, 2001; Derricourt, 2005). The presence of Acheulian bifaces in alluvial river terraces in the southern Negev Desert and at the site of 'Ubeidiya in Israel (around 1.4 Ma) is unequivocal evidence of a Lower Palaeolithic human presence during Lower to Middle Pleistocene times in what are now arid areas (Ginat et al., 2003; Goren-Inbar et al., 2004).

Another possible route was across the narrow and shallow Bab el-Mandeb Strait in the southern Red Sea at times of lower sea level and thence into Arabia and on to Asia, perhaps along the coast (Stringer, 2000). Humans were certainly present on the western side of the Red Sea during the last interglacial (Walter et al., 2000). The Bab el-Mandeb ('Gate of tears' in Arabic) today has a minimum depth of 26 m (Jarosz, 1997) and a minimum width of about 4 km. Using published sea level curves spanning the last 125 ka (Williams et al., 1998, p. 119), it is possible to make a first-order estimate as to when the strait could have been crossed on foot. It could have been dry or very shallow for most of the time between around 80 and 15 ka (although possibly flooded from 65 to 60 ka) and dry again, briefly, at around 115 and 105–90 ka. An exodus at any of these times is broadly consistent with the evidence from both molecular biology and prehistory, neither of which are very precisely dated. More recent work by Lambeck et al. (2011), taking into account isostatic and tectonic factors, and more recent but unpublished bathymetric surveys concluded that although there was never a land bridge, the southern Red Sea in places was sufficiently shallow

and the distances between islands was small enough that quite modest rafts or logs would have enabled a crossing from Africa to Arabia.

A third possible migration route is across the Sahara and then along the eastern shores of the Mediterranean into southern Europe and south-west Asia. Hominid skulls belonging to *H. erectus*/*H. ergaster* have been recovered from Dmanisi in Georgia and have been dated to around 1.75 Ma (Vekua et al., 2002). The presence of *H. erectus* skulls at Zhoukoudian/Choukoutien near Beijing in north-east China and at Solo in Java testify to the ability of these Lower Palaeolithic people to adapt to both temperate regions with very cold winters and to the hot, wet tropics. Given that they had to cross what are today vast and often waterless deserts in order to reach eastern Asia, it seems most likely that such crossings only took place during wetter climatic intervals, when the deserts were able to sustain permanent lakes, wetlands and rivers. By observing flocks of semi-desert birds such as sand grouse (*Pteroclididae*), which fly to permanent water sources to drink at regular times each day, they would have been able to locate secure supplies of water. However, there is no evidence that the Lower Palaeolithic peoples were ever able to occupy deserts during the arid climatic intervals, in contrast to modern desert dwellers, such as the San people of the Kalahari or, until recently, the Walbiri of central Australia. Desmond Clark (1980) used modern ethnographic examples to suggest a model of seasonal movement of small bands of Acheulian hunter-gatherers in North Africa in accordance with the seasonal availability of wild foods, including honey.

The inception of the Middle Palaeolithic dates to around 500,000 years ago in southern Africa and may have been time-transgressive, with many workers claiming an age of around 300 ka for the ESA/MSA transition. The development of heavy-duty choppers and other woodworking tools denotes the increasing use of wood at this time, as does the ability to attach stone spear points to long wooden shafts for hunting larger game. Regional specialisation becomes more evident during the Middle Palaeolithic/MSA (Clark, 1982; Klein, 1989; Van Peer, 1998), including the appearance of tool-making traditions such as the Aterian tanged points that were in common use across the central and northern Sahara during the Late Pleistocene (Van Peer, 1998; Clark et al., 2008). The first entry into Australia some 45,000 years ago was by seafaring people with a stone tool-making tradition that was on the cusp between Middle and Upper Palaeolithic but had unique elements such as edge-ground axes, which date back to at least 25,000 years ago in northern Australia (Schrire, 1982). Elsewhere in Eurasia, polished stone axes are considered to be diagnostic of the Neolithic, showing that it is unwise to be too rigid when using stone typology to determine chronology.

A persistent debate among archaeologists, paleoanthropologists and geneticists concerns the origin of anatomically modern humans (*Homo sapiens*) and whether or not present-day people (*Homo sapiens sapiens*) are descended from one or more widely scattered original human groups (Van Peer, 1998; Underhill et al., 2001;

Mellars, 2006). The preferred hypothesis, although it is not without its critics (e.g., Thorne and Wolpoff, 1981; MacEachern, 2000; Dennell and Roebroeks, 2005), has been named the ‘Out of Africa’ scenario, and is strongly supported by the evidence from mitochondrial DNA (mtDNA), which is inherited through the mother (Cann et al., 1987; Watson et al., 1997; Ingman et al., 2000), and Y chromosome data, which is inherited via the father (Underhill et al., 2001). A comprehensive review by Underhill et al. (2001), supplemented by other studies (Ke et al., 2001; Templeton, 2002; Mellars, 2006), points to multiple episodes of population expansion within Africa and associated migrations of *Homo sapiens* out of Africa into Asia and on to Australia by around 45 ka, as confirmed by the detailed study of an Aboriginal genome (Rasmusson et al., 2011), western Europe by around 45 ka and the Americas by around 13–11 ka. As noted earlier, one route was from north-east Africa and the Sinai via the Levantine corridor (Derricourt, 2005), another was across the Sahara following the last interglacial rivers to the Mediterranean coast (McKenzie, 1993; Rohling et al., 2002; Osborne et al., 2008; Castañeda et al., 2009; Drake et al., 2011; Coulthard et al., 2013). Some of these rivers flowed along the former courses of the Neogene Sahabi rivers of the Sahara which flowed from northern Chad across Libya to the Mediterranean Sea (Griffin, 2002; Griffin, 2006; Griffin, 2011). There is also good evidence from dated lake deposits in the Murzuq Basin of southern Libya that at least four large lakes were present in this now hyper-arid area between 500 and 100 ka, with U-series ages of around 415, 320–300, 285–205 and 138–128 ka (MIS 5e) (Geyh and Thiedig, 2008). These large lakes became progressively smaller during each successive wet phase, indicating progressively less humid interglacial conditions from MIS 11 onwards. Other Middle to Late Pleistocene lakes immediately west of the Saharan Nile have been identified from satellite imagery and were fed in part by overflow from the Nile (Maxwell et al., 2010), providing another possible well-watered route across the eastern Sahara. Yet another potential route was across the southern Red Sea at the Bab el-Mandeb Strait during times of low sea level between short episodes of high interglacial sea levels (Bailey et al., 2007; Armitage et al., 2011; Lambeck et al., 2011).

A recent exodus may have taken place about 70,000 years ago (Ambrose, 1998). If correct, this would have been after the now precisely dated 74 ka huge eruption from Toba volcano in Sumatra, which some workers consider to have been indirectly responsible for a major drop in human population at that time (Ambrose, 1998; Rampino and Ambrose, 2000). Views on the impact of the 74 ka Toba eruption are polarised between those who claim little or no impact (Petraglia et al., 2007) and those who argue for a substantial and adverse impact (Williams et al., 2009a). Williams (2012b; 2012c) summarises this debate and suggests way to achieve better progress for understanding the impact.

Not all movement was solely out of Africa. The mtDNA evidence shows that people moved from the Levantine region of south-west Asia to both North Africa and

Europe around 45–40 ka ago, where they were associated with the development of very distinctive Upper Palaeolithic industries (the Dabban in North Africa and the Aurignacian in Europe) (Olivieri et al., 2006). These conclusions seem secure, but it should always be borne in mind that errors can occur as a result of the ‘founder effect’, which involves the loss of genetic variation that arises when a relatively small number of individuals establish a new population. Refined ^{14}C -dating shows that anatomically modern humans had reached southern Italy by 45–43 ka (Benazzi et al., 2011) and south-west England by 41–39 ka (Higham et al., 2011), bringing new stone tool-making methods with them.

The Last Glacial Maximum (LGM: around 21 ka ago) was a time of extreme global environmental stress. The Y chromosome genetic evidence suggests that during the extreme environment of the LGM, there was a drop in population numbers and people sought refuge in isolated areas (Underhill et al., 2001), from which refugia they branched out as the ice caps and glaciers retreated, temperatures became warmer and conditions improved for *H. sapiens* but not necessarily for the Neanderthals. The evidence from securely dated river and lake deposits, dunes, dust mantles, periglacial and glacial features, plant and animal fossils, marine cores and oxygen isotope studies (all reviewed in earlier chapters), taken together, points to colder global temperatures and greater-than-present aridity between latitudes 30°N and 30°S during and immediately after the LGM. Peak aridity appears to have coincided with early postglacial warming and the release of abundant meltwater into the North Atlantic, which led to a weakening of the oceanic thermohaline circulation system. Precipitation minima and evaporation maxima may have been slightly out of phase with maximum wind intensities, so the dune and dust records need to be interpreted with care. This late Pleistocene phase of aridity was broadly synchronous in both hemispheres and marked the greatest expansion of our deserts since the advent of *Homo sapiens* (and the associated Middle Stone Age cultures) some 200 ka ago. Earlier intervals of major Pleistocene aridity also coincided with global temperature minima. Glacial maxima were times of low atmospheric carbon dioxide content, low atmospheric water vapour content, high ultraviolet radiation, greater windiness, lower temperatures, lower precipitation and increased evapotranspiration, which means that physiological stress to plants and dependent animals would have been great. Given these extreme conditions, it seems probable that prehistoric hominids occupied the great tropical deserts, such as the Sahara and Arabia, during milder climatic intervals, especially the relatively short-lived interglacial phases, when water, plants and animals were more easily available, rather than during the glacial maxima (Williams et al., 1987; Osborne et al., 2008; Castañeda et al., 2009; Drake et al., 2011).

Scattered throughout the Sahara are abundant remains of the stone tools left behind by the Early, Middle and Late Stone Age peoples who once roamed the Sahara during these wetter climatic intervals (Clark, 1980; Gifford-Gonzalez, 2008). The Late Stone Age hunters who preyed on the savanna herbivores were also gifted artists, leaving

behind an enduring legacy of rock engravings and rock paintings depicting the animals they knew so well. With the onset of plant and animal domestication by small groups of Neolithic herders and farmers some 10,000 years ago, the focus of the paintings changed. Cattle camps, showing herds of brindled cattle guarded by men with bows-and-arrows and dogs, were now painted on suitable smooth rock faces in mountainous areas throughout the Sahara (Muzzolini, 1995; Coulson and Campbell, 2001). Some of these paintings show women in their finery, riding oxen just as they do today among the Baggara cattle-owning tribe of western Sudan during their summer migrations north into the desert to search for fresh pasture and to escape the biting *Tabanidae* flies that herald the onset of the rains further south. Others show papyrus or reed canoes similar in design to those still made and used on Lake Tana near the Ethiopian headwaters of the Blue Nile. The plains adjacent to high mountains were preferred occupation sites for these Neolithic pastoralists and their proto-historic successors (Figures 17.5 and 17.6). Small lakes and permanent springs were a guarantee of survival in years when the summer rains failed. The cattle herders ranged as far as Jebel 'Uweinat in south-east Libya, the Tassili sandstone plateaux in southern Algeria and the Air Massif in Niger.

The nature of the interactions between prehistoric peoples and their environment remains the subject of enduring archaeological enquiry. Given the former presence of domesticated cattle, sheep and goats in areas no longer able to sustain them, it is tempting to speculate that they themselves may have accelerated their exodus from the desert. An obvious question to ask is: To what extent did Neolithic overgrazing by large herds of hard-hoofed cattle, sheep and goats accelerate soil erosion by wind and water and initiate humanly induced processes of desertification, especially in the drier second-half of the Holocene? At Adrar Bous in the central Sahara, the rate of sedimentation evident in small valleys around the central granite massif was an order of magnitude greater than it had been during earlier phases of human occupation, suggesting that a combination of grazing pressure allied with progressive (or rapid) climatic desiccation may have caused the increase in erosion, a situation likely to be true of other parts of North Africa at this time (Williams, 1984b; Williams, 1988), as well.

Coastal shell middens provide another important, unambiguous source of information about human occupation of arid lands. For example, the extensive Neolithic and younger shell midden sites along the arid coasts of Mauritania (Petit-Maire, 1979a; Petit-Maire, 1979b) or Peru contain stone tools and other Neolithic artefacts (Sandweiss et al., 2001) and clearly indicate a human presence at the time they were accumulating. In Peru, the shells and fish otoliths (Andrus et al., 2002) have been used to determine changes in the temperature of the seawater along the coast, which also varies between El Niño and La Niña years (see Chapter 23), being warmer in the former and cooler in the latter (Quinn and Neal, 1987). The sudden accumulation of shell middens along this now arid coast some 5,000 years ago has been used as

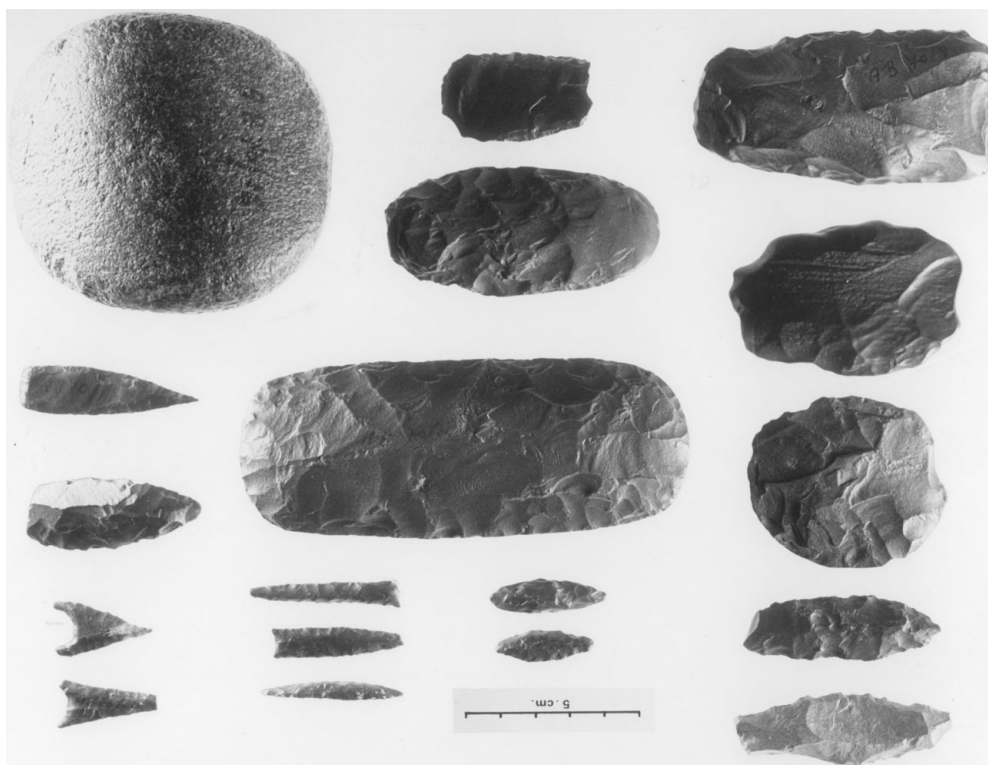


Figure 17.5. Mesolithic and Neolithic stone tools from Adrar Bous, south-central Sahara.

evidence of an episodically wetter climate associated with the inception of a regular El Niño climatic regime (Sandweiss et al., 2001), a theory that is scrutinised in [Chapter 23](#). In certain shell middens along the south-east coast of Australia, there is some evidence of growing pressure on local marine resources, with a decrease in the size of shells collected. Combined with other signs of increasing pressure on local natural resources, such as the consumption of food items from progressively lower trophic levels, local faunal extinctions, skeletal signs of malnutrition and movement into more marginal environments (Cohen, 1977), a case may be made for a late Holocene intensification of prehistoric human occupation in south-east Australia, possibly associated with the arrival of a wave of new immigrants from Indonesia who brought with them the dingo some 5,000 years ago (Mulvaney and Kamminga, 1999).

17.5 Prehistoric butchery sites, fire and faunal extinctions

The inception of stone tool-making appears to be associated with a deliberate increase in meat-eating. The presence of butchery sites in presently arid regions such as the hyper-arid Afar Desert is a useful guide to the former presence of humans in such

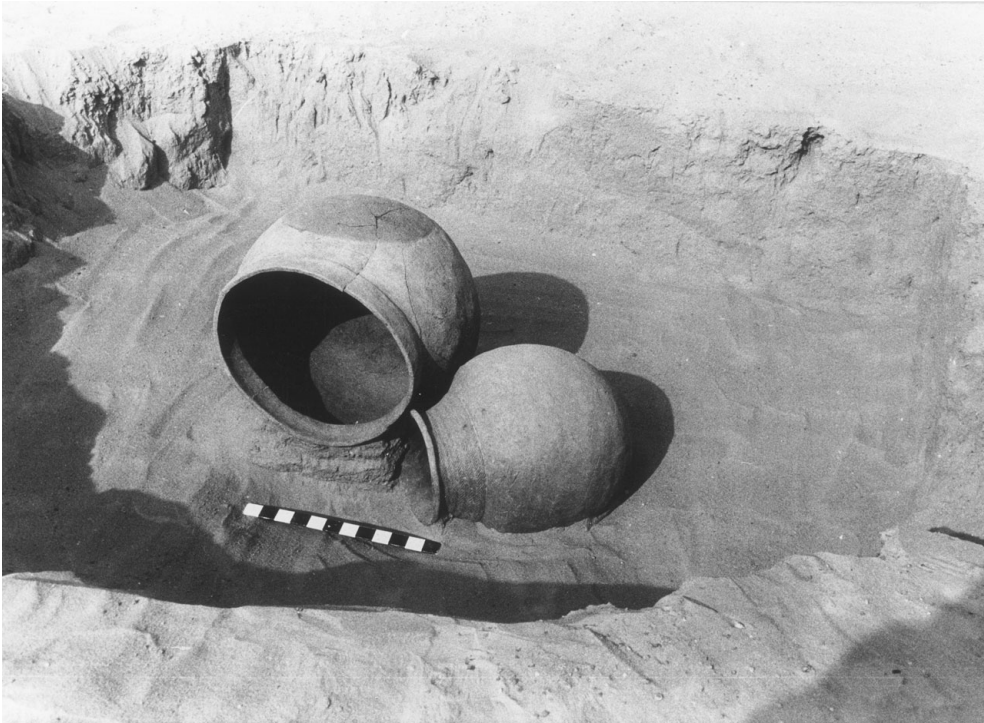


Figure 17.6. Neolithic pots, Adrar Bous, Aïr Mountains, south-central Sahara.

regions (De Heinzelin et al., 2000). However, many sites once claimed to be former butchery sites are no longer considered to be such (Binford, 1981; Brain, 1981b). For a site to be accepted as an unequivocal butchery site, four criteria need to be fulfilled (Binford, 1981). First, the animal carcass must be in primary context. Second, the carcass needs to be associated with the stone tools that were used in the butchery operation. Third, the bones should show signs of cut-marks, preferably oriented at right angles to the long axis of the bones and close to the joints between bones, indicating disarticulation. Finally, the cutting edges of the stone tools should show signs of microwear consistent with cutting through hide and flesh, and, ideally, traces of blood residue. Once these criteria are rigorously applied, it becomes clear that finding a genuine butchery site is a rare event.

In the early stages of human cultural development, our earliest ancestors are probably best described as opportunistic scavengers rather than hunters of big game. They were no doubt capable of killing small animals, just as present-day chimpanzees are occasional hunters of monkeys, but the only initial advantages they had when it came to obtaining extra protein in the form of meat were the possession of sharp stone flakes for cutting through the hides of large, dead animals and the ability to operate during the heat of the African day, thanks to their ability to remain cool by sweating.

The big cats of the African savanna cannot cool themselves in this way and therefore doze during the day and hunt at night, when it is cooler. It is only much later, with the safety provided by long-distance spear throwing, that our Middle Palaeolithic forebears became more proficient at hunting larger animals, and by Upper Palaeolithic and Mesolithic times, the use of pit traps and other snares became common. High in the semi-arid Kaimur Ranges of the Vindhyan Hills in north-central India, north of the Middle Son Valley, Mesolithic rock shelters show paintings of pit traps in which a now extinct Indian rhinoceros is caught. The Bega hill people in this region today remain hunter-gatherers, and they are adept at making fire using just a wooden base with a slight hollow in it to position kindling and a thin straight stick of hard wood placed vertically in the hollow and rotated rapidly between the palms of the hands. Flames ensue in several minutes.

17.5.1 Late Pleistocene megafaunal extinctions in Australia

The Aboriginal people of seasonally wet tropical northern Australia inhabit a land that is a rainless desert for more than half the year. They have long used fire as a hunting tool and can gauge when and where to light the fire so that it remains under control, using their local knowledge of when the winds will change direction during the day and blow the fire back onto burnt land, which acts as a fire-break (Haynes, 1991). So widespread was the practice of burning the land throughout Australia, including Tasmania, at the time of European contact that the archaeologist Rhys Jones (1968) coined the phrase ‘firestick farming’, although many would cavil at this exuberant use of the word ‘farming’, which normally connotes cultivation of the land for growing food crops. Be that as it may, there is no doubt that the human use of fire would have greatly facilitated hunting in a variety of ways, including stampeding animals towards the hunters, removing the tall, dead grass and facilitating the growth of palatable, new green grass at the start of the wet season. This brings us to a hotly debated set of questions. Were the widespread and well-documented extinctions of the larger animals in the Americas, Eurasia and Australia towards the end of the Pleistocene a result of climate change, human predation, human modification of the environment through burning or some other natural catastrophe, such as meteorite impact or volcanic eruption? Needless to say, any of these factors may have operated at any one time and place, either individually or in combination with one or more of the others.

Some of the oldest evidence for Quaternary animal extinctions comes from Australia, where 90 per cent of the larger kangaroos, together with other giant species, had vanished by around 45,000 years ago, shortly after the time that humans first occupied the continent. However, it is curious that although hunting is claimed as a major cause of the demise of the Australian megafauna, once humans arrived on that continent some 50,000 years ago (Miller et al., 1999; Roberts et al., 2001; Gillespie, 2008),

not a single convincing butchery site has yet been identified, in strong contrast to the abundance of Moa butchery sites in New Zealand (Martin, 1984, pp. 391–392) and of early Holocene bison and mammoth butchery sites in North America (Martin, 1984).

Arguments put forward to account for the lack of early Australian butchery sites refer vaguely to past climatic changes or to changes in river behaviour. For example, Flannery (1994, pp. 199–200) refers to erosion, ‘the restless Earth’ and ‘the dynamic nature of the Earth’ to account for the lack of evidence ‘relating to how humans affected Australia’s giant marsupials, birds and reptiles’ (op. cit., p. 199). This reasoning is fair enough, perhaps, but it does not explain why butchery sites survive so well in Africa from Acheulian times onwards, having successfully endured repeated environmental changes during that time (Clark, 1982; Isaac, 1982). It might be simpler to seek other and additional causes for the demise of the large marsupials, such as the progressive desiccation of Australia that set in around 50 ka ago (Cohen et al., 2010b), which was also about the time when humans first moved into Australia (Bowler et al., 2003). Although it is a truism that absence of evidence is not evidence of absence, the lack of true butchery sites associated with Australian megafaunal remains may simply indicate that large marsupials were not hunted to extinction but died out from other causes, such as habitat modification caused by humans burning the vegetation to assist in hunting.

In order to test these hypotheses, Prideaux et al. (2007; 2009; 2010) have conducted a series of elegant and exemplary studies. A cave known forbiddingly as Tight Entrance Cave in south-western Australia contains the ‘richest and most diverse assemblage of Late Pleistocene vertebrates known from the western two-thirds of Australia’ and is also the ‘only site on Earth known to have sampled a mammal community for 100 ka preceding regional human arrival and then subsequently’ (Prideaux et al., 2010, p. 22157). Analysis of the stable carbon and oxygen isotopic composition of land snail shells composed of aragonite (which therefore have not undergone any recrystallization or secondary diagenesis) demonstrates significant environmental (and climatic) change from 70 ka onwards, culminating in the extreme cold temperatures and aridity of the LGM at around 21 ka. However, the extinctions predate this later extreme event by some 20,000 years. Counts of coarse and fine charcoal washed into the cave denote local and regional fire frequency, respectively. The earliest evidence of a human presence at this site is dated to 49 ± 2 ka, while the most recent age for the local presence of an extinct mammal is 40 ± 2 ka, showing that the extinctions were progressive and not the abrupt events posited by Paul Martin in his North American Blitzkrieg model of hunting-induced extinctions (Martin, 1967). Prideaux et al. (2010) conclude that ‘on balance, human impacts (e.g., hunting, habitat alteration) were most likely the primary driver of the extinctions’, but they go on to caution that ‘it is equally probable that the ultimate extinction “cause” was complex, and that landscape burning and increasing aridity helped fuel the extinction process’ (op. cit., p. 22161).

As Williams et al. (1998) also noted in their earlier discussion of prehistoric extinctions, unicausal explanations are seldom persuasive.

Some workers have pointed out that much of the evidence for faunal extinctions come from the better-watered south-western and south-eastern periphery of the continent and not from the more arid interior, although this stricture does not apply to the extinction of the giant flightless bird *Genyornis* in and around the Lake Eyre Basin that was so well-documented by Miller and his co-workers (Miller et al., 1999). In response to this criticism, Prideaux et al. (2009) completed a detailed investigation of what the largest-ever kangaroo, *Procoptodon goliah*, was actually eating in the arid zone of south-east Australia. From the carbon isotopic composition of the tooth enamel, combined with tooth microwear patterns, they concluded that *P. goliah* was a specialist chenopod browser, most likely of the saltbush genus *Atriplex*, and that it drank more water in the arid zone than its grazing contemporaries did, much like sheep that eat saltbush today. Because saltbushes and chenopod shrublands do not burn easily, they discounted fire as a major cause of the demise of *P. goliah*, concluding that it had survived many previous climatic cycles from more to less arid conditions, so hunting of this tall, conspicuous animal by the first Australians remained the most plausible cause. However, this is an inference by default, because no positive evidence of actual hunting is forthcoming.

It is useful to recall that most of the larger marsupials that became extinct in the late Quaternary were primarily browsers (eaters of shrubs and leaves from trees) rather than grazers (grass-eaters). The long-term change in the Australian flora took place against a backdrop of rapid climatic fluctuations superimposed on long-term desiccation (Martin, 2006). The result as far as the browsers were concerned would have been a progressive impoverishment of their habitat during the Quaternary, with grassland expanding at the expense of forest and woodland. The arrival of humans, allied with their initial ignorance of how best to control the destructive impact of fire, would have aggravated the pressure on the large browsers with their slow breeding cycles. As the large marsupials declined in number, so too did their large carnivorous predators, such as *Thylacoleo* and *Megalanina*. In the words of John Calaby, the late Pleistocene arrival of humans on the Australian landscape was simply the last nail in the coffin for a group of animals already on the way to extinction (Calaby, 1976), a view accepted by Williams et al. (1993; 1998).

Another approach to this problem is to investigate the possible consequences (as opposed to the causes) of late Quaternary extinctions of the larger herbivores on local and regional habitats. Flannery (1994) suggested that the late Quaternary extinction of the large marsupials in Australia would have altered the vegetation, effecting a change in fire regime and causing a major ecosystem change. Miller et al. (2005) also argued for Australian 'ecosystem collapse' shortly after the demise of the Pleistocene megafauna, which according to their work on the now extinct, giant flightless bird *Genyornis*, vanished soon after the arrival of humans (and humanly lit bushfires) on

that continent. The inferred outcome in arid central Australia was a change from a mosaic of trees, shrubs and grasslands adapted to drought to the modern fire-adapted desert scrub.

Lopes dos Santos et al. (2013) sought to test these hypotheses by examining the possible impact of extinction of the large marsupial fauna on the ecosystems of the Murray-Darling Basin, which drains about one-fifth of the Australian continent. She and her co-workers analysed a 140 ka marine sediment core obtained from a canyon several hundred kilometres off the mouth of the Murray River in south-east Australia. They used changes in accumulation rates of levoglucosan as a proxy for changes in biomass burning and changes in the stable carbon isotopic composition of higher plant wax *n*-alkanes as a measure of vegetation change inferred from changes in the proportions of plants following the C₃ and C₄ photosynthetic pathways, as deduced from the stable carbon isotopic record (see Chapter 7). The age control on the core was of sufficient resolution to enable them to test the environmental changes in the Murray-Darling Basin before and after the putative extinction of the Australian megafauna between 48.9 and 43.6 ka. They found proxy evidence of an abrupt and short-term decrease in the abundance of C₄ vegetation at around 43 ka, followed by an increase in biomass burning lasting about 3,000 years. They concluded that because these two events occurred after what they term the ‘main period of human arrival’ and megafaunal extinction in Australia, the change in vegetation cannot have caused the extinction but may have been a consequence of it, as initially hypothesised by Flannery.

17.5.2 Late Pleistocene megafaunal extinctions in arid North America

In few parts of the desert world have late Pleistocene faunal extinctions received as much scholarly attention as in North America, where the relatively late arrival of sophisticated hunters (and makers of the Clovis spear points) is thought to have brought about the rapid demise of the larger mammals (Martin and Wright, 1967; Martin, 1984; Martin and Klein, 1984). The apparent speed of the human occupation of North America, as well as that of the faunal extinctions, prompted Martin (1967) to invoke the World War II German military metaphor for invasion and warfare at lightning speed (*Blitzkrieg*) as a description of the process. Although seductive in its simplicity, this interpretation has been rejected or at least modified by those who point out that the terminal Pleistocene was also a time of rapid climatic change in North America (and, indeed, elsewhere), with substantial modification of plant communities and a continent-wide repatterning of food supplies.

Several recent studies have cast doubt on the *Blitzkrieg* model. Haile et al. (2009) investigated ancient DNA and concluded that mammoths and horses survived in interior Alaska until at least 10.5 ka, or thousands of years after the initial arrival of humans in North America. Lorenzen et al. (2011) summarised studies of ancient

DNA supported by 1,439 directly dated megafaunal sites and 6,291 radiocarbon ages associated with the Upper Palaeolithic human presence in Eurasia, and they found that each species responded differently to the effects of climate change. They concluded that climate change alone explains the extinction of the Eurasian musk ox and woolly rhinoceros but that a combination of climatic and human impacts seems to be responsible for the demise of the wild horse and steppe bison.

Polyak et al. (2012) bring a novel approach to this problem by using the $\delta^{13}\text{C}$ and $\delta^{234}\text{U}$ values in speleothem calcite from Fort Stanton Cave in southern New Mexico as a proxy for effective precipitation, supplemented by ages obtained from rim pools in the Big Room of Carlsbad Cavern (see also [Chapter 14](#)). They found that a very severe drought followed a moist pluvial interval and afflicted the south-western United States from just before 14.5 until 12.9 ka or soon thereafter, an interval of time that is broadly synchronous with the 14.6–12.8 ka Bølling/Allerød warming event evident in the Greenland ice core record. They noted that the last appearance of sixteen out of thirty-five mammal genera that became extinct between 13.8 and 11.4 ka overlapped with this 1,500 year drought in the arid south-west and predated both the arrival of the Clovis hunters and the highly controverted cometary impact invoked (most aptly) by Firestone et al. (2007) as possible causes of the extinction of North American large mammals. They dismissed evidence of a sparse pre-Clovis human presence in North America (Waters and Stafford, 2007) as unlikely to have a significant impact, although they did not rule out some human contribution to Pleistocene extinction. Once again, the question of causes remains open. Two difficulties noted by Martin (1984) in his global overview of the ‘prehistoric overkill’ model remain as valid today as they were thirty years ago: ‘There is no guarantee that the time of extinction will inevitably be found by archaeologists’ (op. cit., p. 392), and ‘a conceptual difficulty has centered on the failure of the fossil record of many regions to disclose ample evidence of extinct faunas in kill sites in any other cultural context’ (op. cit., p. 396).

17.6 Use of prehistoric stone tools as stratigraphic markers

Prehistoric archaeology can contribute to our knowledge of climatic change in deserts in one additional albeit indirect way, namely, by providing information akin to that given by more orthodox forms of biostratigraphic markers, such as plant and animal fossil assemblages (see [Chapters 3 and 16](#)). For example, in localities such as Olduvai Gorge in semi-arid Tanzania, which has a long and reasonably continuous record of prehistoric human occupation, the progressive changes in stone tool assemblages over time have been grouped into Early Stone Age (ESA – Oldowan, followed by the Acheulian), Middle Stone Age (MSA), Late Stone Age (LSA) and Neolithic. Potassium-argon dating of volcanic ash or welded tuff units located above and beneath the stone tool-bearing horizons has allowed the older African stone tool-making traditions to be reasonably well dated. In the case of the younger assemblages, radiocarbon dating

has yielded reliable ages back to about 35,000 years ago, and in the last few years, luminescence dating has been applied to sediments several hundred thousand years in age. As a broad generalisation, the ESA in Africa began around 2.5 Ma ago, with the Acheulian appearing around 1.5 Ma ago, together with the first signs of fire use; the MSA is bracketed between about 500/300 and 50 ka, and the LSA dates to between about 50 to about 11 ka, after which early plant and animal domestication denotes the inception of the Neolithic.

In desert areas devoid of volcanic ash and other sediments suitable for radiometric dating, such as the Sahara, Kalahari and Namib deserts, the ages obtained elsewhere for stone tool assemblages have been used to establish a chronology for environmental changes evident in the depositional record. This approach was used with some success at Adrar Bous, a small mountain situated in the heart of the Sahara, some 1,500 km from the nearest coast (Williams, 1987; Williams, 2008) and in the piedmont west and south of Jebel Marra in Darfur, western Sudan (Williams et al., 1980; Philibert et al., 2010).

In the latter region, which is semi-arid today, the botanist Gerald Wickens (1975a, 1975b; 1976a; 1976b) found leaf fossils of *Combretum* and the oil palm *Elaeis guineensis* in reworked volcanic tuffs near the village of Umm Mari between Kas and Nyertete townships (Chapter 15, Figure 15.5). The oil palm shows the former presence of tropical rainforest. Wickens believed that the fossils were probably early Holocene in age and suggested that I revisit the site to check this, which I was able to do in January 1976, when I found a Developed Oldowan/Early Acheulian stone tool assemblage comprising fresh basalt choppers immediately beneath the fossil-bearing tuff, together with bifacial and unifacial choppers, push-planes, discoids, hammer-stones and flake scrapers (Williams et al., 1980). Similar assemblages in East Africa span a maximum time range from 1.5 to 0.3 Ma and a more probable time range of 1.2 to 0.8 Ma (Clark and Kurashina, 1979; Williams et al., 1979; Isaac, 1982; Gowlett, 1984; Clark, 1987; Owen et al., 2008).

Some 90 km north-east of Umm Mari near the village of Barbis, more than 5 m of finely laminated diatomites testify to the former presence of a deep freshwater lake (Philibert et al., 2010). A thin layer of sandy alluvium with sporadic basalt and trachyte gravels overlies the diatomite. Among the gravels were occasional large-trimmed flakes, flake scrapers, bifacially worked choppers and scrapers, as well as the broken butt of an Acheulian hand-axe. Some of the artefacts were slightly abraded, one was very fresh with sharp edges and three were heavily abraded. According to Dr John Gowlett, who examined the collection, typologically similar artefacts occur at Olorgesailie in Kenya and Olduvai Bed IV in Tanzania and straddle the Brunhes-Matuyama paleomagnetic boundary (0.78 Ma), so an age range of around 0.8 ± 0.3 Ma is likely for the stone tool assemblage. Later work has confirmed this age estimate (Isaac, 1982; Gowlett, 1984; Owen et al., 2008). The evidence is circumstantial, but it is likely that the hominids who made the stone tools now found on the surface of the

diatomite were living close to the lake about 0.8 ± 0.3 Ma ago and that once the lake dried up, some of the stone tools were washed onto the now dry surface of the lake during occasional sheet-floods. Elsewhere in northern and eastern Africa, Acheulian occupation sites are invariably associated with the presence of freshwater lakes, rivers or springs (Clark, 1980), suggesting a close dependence on permanent sources of water.

17.7 Conclusion

Prehistoric evidence of a former human presence in areas that are now too arid to support much life is generally associated with other evidence from lake and river sediments, fossil pollen grains and macrofossil remains that show that conditions were wetter at those times. The archaeological record has the double advantage of providing skeletal remains (that can be dated and analysed using stable isotopes to determine past diet in the case of Holocene remains), as well as associated stone tool assemblages. Determining the functions of prehistoric tools is not always easy, and current systems of stone tool classification are based on size, shape and overall morphology, rather than function, although studies of microwear and residues left on the cutting edges of certain tools are throwing some light on prehistoric tool-use and function. In the absence of other means of dating, both stone tool assemblages and skeletal remains can be used to provide a relative chronology of the deposits in which they occur.

The late Miocene and Pliocene hominids are confined to Africa and are now well dated. The oldest known stone tools come from the Gona tributary valley to the Middle Awash Valley of the Ethiopian Afar Rift and date to around 2.5 Ma. These pebble tools, or Oldowan tools, were used for more than 1 million years before being succeeded by Acheulian hand-axes and cleavers. The Oldowan and Acheulian together comprise the Early Stone Age, or Lower Palaeolithic. Oldowan and Acheulian tools are found across Eurasia, and indicate periodic movement out of Africa by *Homo erectus*/*Homo ergaster* via the Levantine Corridor during times when the climate was wetter. The Middle Palaeolithic/Middle Stone Age began about 500 ka in southern Africa but may not have originated until 300 ka further north. One hallmark of the Middle Stone Age was the use of hafting stone projectile points made from previously shaped cores. Soon after 50 ka, the Middle Stone Age gave way to the Later Stone Age, or Upper Palaeolithic. This was a time of great regional diversity in stone tool-making and of the proliferation in certain localities of rock art, sculpture and the manufacture of stone artefacts of great beauty, no doubt used for trade and for establishing social networks.

A contentious and still unresolved issue concerns the possible role of prehistoric humans in faunal extinctions, with some evidence pointing to climate change as a key factor, some suggesting over-predation from hunting and some indicating habitat

modification from the human use of fire. Another formerly much-debated question concerns the causes of plant and animal domestication in widely separated regions soon after 11 ka and the start of the Neolithic. Certainly, the development of appropriate tools for harvesting, grinding and storing cereal grains is an obvious prerequisite, but other less easily discerned social and economic factors no doubt played a role, as did stress on local natural resources from a growing population (Cohen, [1977](#)).

The use of genetic evidence, notably mitochondrial DNA inherited from the mother and Y chromosome data inherited via the father has proven to be a powerful means of determining when and whence human populations migrated out of Africa into Eurasia, Australia and the Americas.

18

African and Arabian deserts

Il serait cependant imprudent d'imaginer que l'heure des synthèses véritables a déjà sonné: le Sahara est d'une ampleur océanique, les lieux sérieusement étudiés y demeurent punctiformes, il subsiste encore nombre d'incertitudes, notamment dans le détail des chronologies paléoclimatiques ou des stratigraphies du Pléistocène. Il faut l'avouer: la part de l'hypothèse reste nécessairement très considérable, on doit honnêtement le reconnaître.

Nevertheless, we would be unwise to delude ourselves that the time is ripe for a definitive account. The Sahara is as large as the ocean, and the sites studied in depth remain mere pinpricks upon its surface. Many issues remain unresolved, notably the detailed chronology of climatic changes and of Pleistocene stratigraphic sequences.

The working hypothesis still has a major role to play, and honesty requires us to recognise this fact.

Théodore Monod (1902–2000)
The Sahara and the Nile
(1980, foreword, pp. xiii, xv)

18.1 Introduction

The aim of this chapter is to provide a synthesis of the Cenozoic and, in particular, the Quaternary environments of the deserts of Africa and peninsular Arabia, including the Kalahari, Namib, Sahara, Afar, East African Rift, Sinai, Negev, Arabia, Yemen and Oman. Much of the evidence has already been reviewed in the earlier specialist chapters but in a fragmentary fashion, so an overview is warranted. Because the tropical northern deserts are more or less contiguous, it is appropriate to treat them together while noting any important differences. We conclude with the southern African deserts.

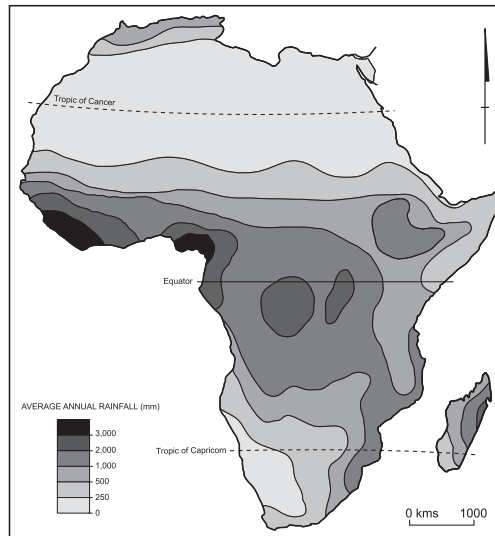


Figure 18.1. Mean annual precipitation in Africa. (After *The Times Atlas of Africa*, 2010.)

18.2 Present environment

The major elements of the present environments in this vast region, including climate and topography, have been described in detail elsewhere (Griffiths, 1972; Rognon and Williams, 1977; Tyson, 1986; Hastenrath, 1991; Nicholson, 1996; Gasse et al., 2008; Nicholson, 2011). So too have the fundamental causes of aridity (Chapter 2). Only a brief summary is therefore needed here. The topography of Africa and Arabia plays an important role in regional climate. The great escarpments of southern Africa and southern Arabia-Yemen-Oman and the highlands of Ethiopia, Kenya and Uganda generate orographic rain and create rain shadows on their leeward sides, accentuating aridity in the lowlands downwind (Figure 18.1). Seasonal migration of wind and pressure belts (Figure 18.2), especially the Intertropical Convergence Zone (ITCZ) in the seasonally wet tropics and the westerlies in the winter rainfall regions of south-west and north-west Africa, determine the amount and pattern of summer and winter precipitation in both northern and southern deserts and their margins. As a very broad generalisation, the western half of Africa receives most of its rain from the Atlantic and the eastern half from the Indian Ocean, although certain areas, such as the Ugandan headwaters of the White Nile, receive rain from both sources. The southern Sahara receives summer rainfall from the Atlantic in the west and from the Indian Ocean in the east. There are occasional inputs from winter depressions, such as along the Atlantic coast of the Sahara and along the Red Sea. These can bring exceptionally heavy rainfall, as in the central Mauritanian desert in January 2004 and in the northern Red Sea and Dead Sea Rift in November 2009.

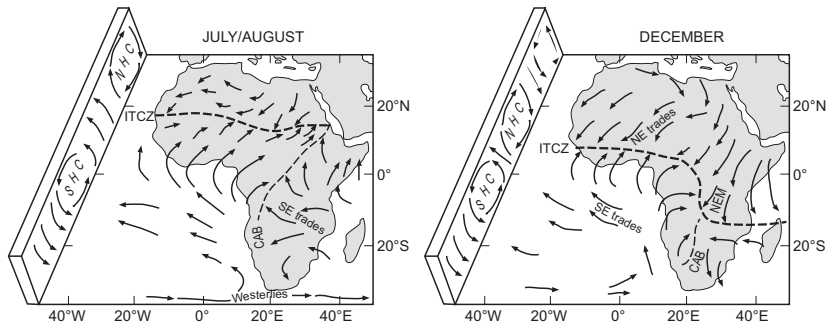


Figure 18.2. Surface winds and frontal locations (a) during July and August and (b) during December. (Modified after Nicholson, 1996, and Gasse et al., 2008.) ITCZ is Intertropical Convergence Zone; CAB, Congo Air Boundary; NEM, northerly East African monsoon. In section: SHC, southern Hadley Cell; NHC, northern Hadley Cell.

Aridity over Arabia and North Africa is accentuated by the subtropical easterly jet stream. Past changes in the strength, sinuosity and latitudinal position of the jet stream would have had important climatic repercussions for both regions (Rognon and Williams, 1977). In southern Africa, the seasonal fluctuations in Antarctic pack ice have a strong influence on the latitudinal displacement of the winter westerlies, which would have brought more rain to south-west Africa during times of maximum ice extent in Antarctica (Stuut et al., 2004; Chase and Meadows, 2007; Gasse et al., 2008). In peninsular Arabia, the south-west monsoon brings rain to the coastal fringe and adjacent uplands in summer. In winter, the dominant wind is the *Shamal* (Arabic for ‘north’), which flows eastwards along the north coast of the Sahara across the Sinai and southern Negev before curving in a clockwise direction to the south and south-west (Chapter 8, Figure 8.9). The Shamal is a dry wind by the time it blows to the south, mobilising sand and causing dust storms.

18.3 Cenozoic desiccation of the Sahara and adjacent deserts

The Sahara is the largest hot desert in the world. It extends more than 4,800 km, from the Atlantic coast of Mauritania in the west to the arid Red Sea Hills in the east, and is continued eastwards across the deserts of Arabia, Iran, Afghanistan and Pakistan to the Thar Desert of Rajasthan in north-west India, a total distance at the Tropic of Cancer of about 9,500 km. The northern limits of the Sahara coincide with the southern margins of the Atlas Mountains in north-west Africa and merge eastwards into the Sinai and Negev deserts. The southern limit of the Sahara is more diffuse and has been defined by some French geographers as the northern limit of the spiny *cram-cram* grass (*Cenchrus biflorus*), a bane to the traveller on foot but a boon in times of extreme drought when famine threatens.

Given its proximity to Europe and its aura of mystery, the Sahara has long exerted an attraction for hardy travellers from that continent. Scientific explorers from many nations have contributed to our knowledge of past and present environments in the Sahara, but those from France in particular have made an outstanding contribution to our understanding of past climates in this vast region ([Chapter 5](#)). This work ranges from the heroic era of more or less solitary exploration and mapping (Bordet, 1952; Monod, 1958; Coque, 1962; Faure, 1962; Chavaillon, 1964; Williams, 1966; Rognon, 1967; Conrad, 1969; Black and Girod, 1970; Fabre, 1974) to the intensive multidisciplinary teamwork of today (Beuf et al., 1971; Fontes et al., 1983; Fontes et al., 1985; Pachur et al., 1990; Hoelzmann et al., 2004; Osborne et al., 2008; Williams et al., 2010b; Drake et al., 2011).

It is appropriate to consider the Sahara, Afar and Arabian deserts together, because for much of their geological history, they were part of the same continent and subject to the same Mesozoic and early Cenozoic northward lithospheric plate movement. Their history only began to diverge as a result of late Cenozoic uplift, rifting and the creation of the Red Sea and Gulf of Aden. The Cenozoic (65 Ma to present) includes the Quaternary (2.6 Ma to present), and the Quaternary includes the Pleistocene (2.6 Ma to 11.7 ka) and the Holocene (11.7 ka to present).

During the Triassic, Africa was part of the Gondwana supercontinent, together with South America, Antarctica, Australia and India. The separation of Gondwana into the two continents of West Gondwana (Africa and South America) and East Gondwana (Australia, Antarctica and India) was accomplished during the Jurassic, although it had begun before then. Further break-up of these two huge continents took place during the Cretaceous ([Chapter 3](#), [Figure 3.2](#)). In Africa, the Cretaceous equator ran diagonally across the Sahara from southern Nigeria through central Chad, northern Sudan and Egypt into Arabia. The northward movement of Africa during the late Mesozoic and Cenozoic resulted in a southward shift of the equatorial rainforest and led to the long-term desiccation of North Africa as it moved into tropical latitudes characterised by dry subsiding air. The slight clockwise rotation of the African plate during the Miocene and Pliocene brought it into contact with the European plate (Habicht, 1979; Owen, 1983; Williams et al., 2004). The ensuing crustal deformation in North Africa led to uplift of the Atlas Mountains in the north-west and was roughly synchronous with volcanism and uplift of the Hoggar, Tibesti, Air and Jebel Marra uplands, creating the major elements of the present-day topography ([Figure 18.3](#)). The onset of aridity was not synchronous across the Sahara; it began earlier in Morocco, Algeria and Tunisia than it did in Egypt and Sudan, as shown by the abundance of Mesozoic and younger evaporite formations in the north-west Sahara, which by then had already reached dry tropical latitudes (Coque, 1962; Conrad, 1969; Williams 1984a).

Cenozoic volcanism and tectonism contributed to the desiccation effected by the northward movement of the African plate, bringing North Africa into dry subtropical

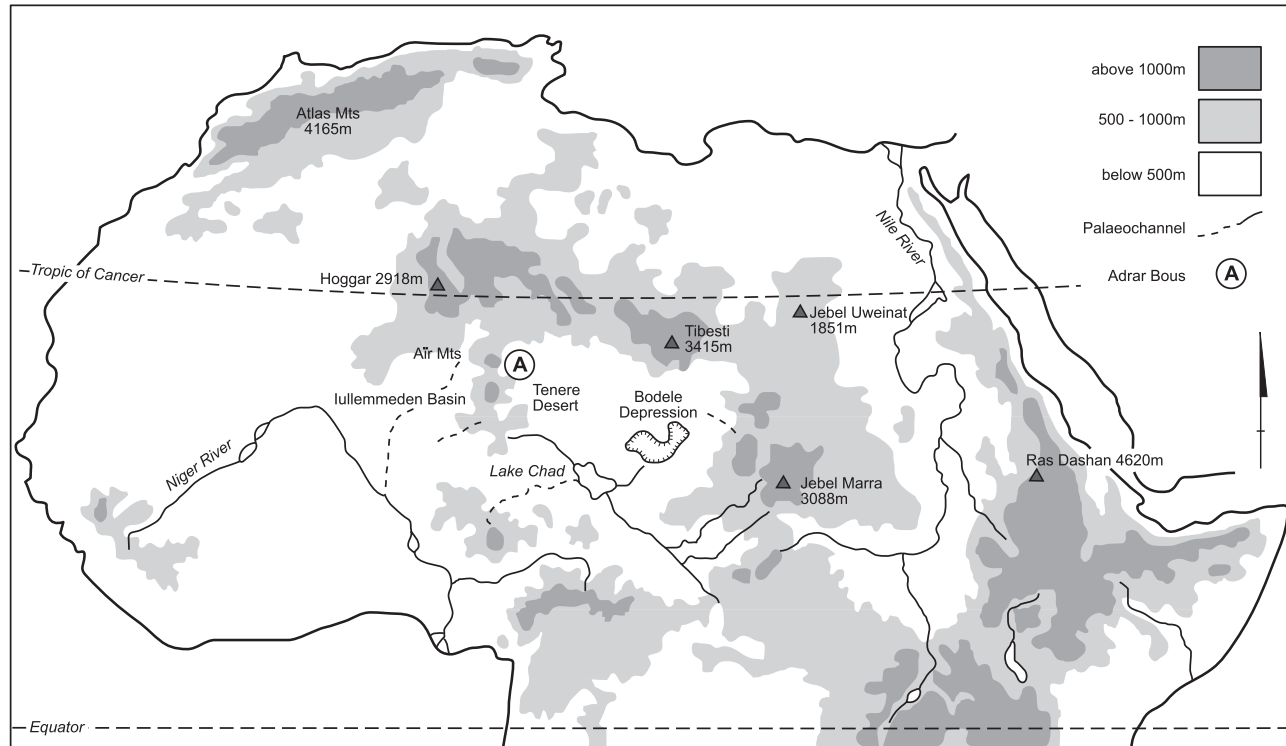


Figure 18.3. Saharan uplands. (A) denotes Adrar Bous massif in the geographical heart of the Sahara.

latitudes. The Miocene and later volcanism in the central and southern Sahara that created the high mountains of the Hoggar, Tibesti and Jebel Marra was preceded by prolonged deep weathering, during which kaolinitic and bauxitic weathering profiles up to 45 m thick developed on rocks of Eocene to Precambrian age along the southern Sahara (Greigert and Pognet, 1967; Williams, 2009a). Uplift in the mid-Cenozoic resulted in a change from chemical sedimentation to dominantly clastic sedimentation in what is now the central and southern Sahara (Faure, 1962; Greigert and Pognet, 1967, p. 157). Rejuvenated rivers flowing down from the great watersheds of Tibesti, the Hoggar and the Air deposited around the uplands the fluvial gravels, sands and clays known to French geologists as the *Continental terminal* (Faure, 1962; Greigert and Pognet, 1967).

Very large, dry, sinuous river valleys are clearly visible on the early Gemini space photographs of Libya and Tibesti (Pesce, 1968) and on the later satellite imagery of the eastern Sahara (Griffin, 1999; Griffin, 2002; Griffin, 2006). One such valley can be traced north from the Chad Basin into the hyper-arid desert of south-east Libya, where it is bounded by the Tibesti volcanic massif to the west and three highly dissected Nubian sandstone plateaux to the east (Williams and Hall, 1965; Pesce, 1968). Griffin (1999; 2002; 2006) has analysed these valleys in detail. He reviewed the sedimentary evidence from marine and terrestrial sites in and around the Mediterranean, the Gulf of Suez and the Red Sea and concluded that the Messinian Zeit Formation (7.04–5.34 Ma) was laid down during a time of high rainfall and high fluvial sediment yield, reaching peak monsoonal activity in the Late Messinian (Griffin, 1999), at a time when the Straits of Gibraltar were closed and the Mediterranean had dried out and become a salt desert. This occurred during the Messinian Salinity Crisis, which is now precisely dated as 5.96–5.33 Ma (Cosentino et al., 2013). Griffin called these Late Neogene rivers the Sahabi Rivers and deduced that they continued to cross central North Africa until about 4.6 Ma ago, when the monsoon shifted to the south and the eastern Sahara became drier (Griffin, 2002).

Post-Eocene uplift in the Sahara triggered a widespread phase of late Cenozoic erosion within major massifs such as the Tibesti and the Hoggar, as well as in more isolated ring complexes such as Jebel Arkenu and Jebel ‘Uweinat in south-east Libya or Adrar Bous (A in Figure 18.3) in central Niger. The mid-Cenozoic drainage system in North Africa appears to have been a highly efficient and well-integrated system which kept pace with the various epeirogenic uplifts across the Sahara.

The Nile cut down through the Nubian Sandstone capping the Sabaloka ring complex to form the Sabaloka Gorge north of Khartoum – one of the many instances of superimposed Cenozoic drainage in the Sahara (Grove, 1980; Williams and Williams, 1980; Thurmond et al., 2004). The early Cenozoic mantle of deeply weathered rock was almost entirely removed from the uplands of the southern Sahara, leaving a bare and rugged landscape of gaunt rocky pinnacles and boulder-mantled slopes. Episodic deep weathering, followed by episodic erosion and the exhumation of the



Figure 18.4. Granite boulders exhumed from a deep weathering profile, Adrar Bous, south-central Sahara.

weathering front (Figure 18.4), became the geomorphic norm of the later Cenozoic (Dresch, 1959; Thorp, 1969; Williams, 1971). There seems little doubt that Neogene tectonic movements performed a dominant role in the initial pulse of erosion, but the Quaternary climatic oscillations became increasingly important erosional pacemakers thereafter (Williams et al., 1987).

The sandy colluvial–alluvial debris eroded from the Saharan uplands was carried away from the mountains by the Neogene and Early Quaternary rivers to be in part deposited in late Cenozoic marine deltas such as those of the Nile, the Niger and the Senegal. However, a considerable proportion of the sediment began to accumulate in the closed interior basins created during the course of late Mesozoic and Cenozoic faulting, rifting and epeirogenic movements. It was the unconsolidated Neogene sediments laid down in large subsiding sedimentary basins, such as the Kufra-Sirte Basin in Libya and the Chad Basin, which provided the source material for the late Pliocene and Quaternary desert dunes (Chapter 8). Miocene tectonic uplift in East Africa may have contributed to the desiccation in this region from about 8 Ma onwards (Sepulchre et al., 2006). In the Chad Basin, Servant (1973) identified wind-blown sands in a number of very late Cenozoic stratigraphic sections. He concluded that the onset of aridity and the first appearance of desert dunes in this part of the southern Sahara was a late Tertiary (i.e., pre-Quaternary) phenomenon. Using fossil and sedimentary evidence, Schuster et al. (2006) have since demonstrated that the

onset of recurrent desert conditions in the Chad Basin began at least 7 Ma ago. Further north, in the Hoggar, some elements of the Late Tertiary flora were already physiologically well-adapted to aridity (Maley, 1980; Maley, 1981; Maley, 1996). If we accept the sedimentological evidence of Servant (1973) and of Schuster et al. (2006) and the palynological evidence of Maley (1996), then it follows that the onset of climatic desiccation and the ensuing disruption of the integrated Neogene Saharan drainage network (Griffin, 2006) was a feature of the very late Cenozoic but long pre-dated the arrival of *Homo sapiens*.

The late Miocene and early Pliocene climate in the Chad Basin fluctuated repeatedly, favouring animals adapted to highly varied ecosystems, including lake, lake margin, riparian, woodland and savanna habitats (Griffin, 2006). It was within this varied set of habitats that the late Miocene Toumaï hominid *Sahelanthropus tchadensis* (TM 266) emerged (Brunet et al., 2005; see Chapter 17). As we saw in Chapter 10, McCauley et al. (1982; 1986) and McHugh et al. (1988; 1989) used shuttle-imaging radar to identify a series of ancient river valleys in the eastern Sahara, some of them former tributaries of the Nile. These valleys range in age from Miocene to Quaternary, with the younger channels associated with Acheulian and more recent artefact assemblages (Chapter 17). Small rivers occupied many of the Neogene river valleys during wetter intervals in the Quaternary (Pachur and Altmann, 2006; Osborne et al., 2008; Drake et al., 2011), but they were never as large as the Neogene and older river systems, although they would have allowed the passage of plants, animals and small bands of humans from central to northern Africa.

Williams et al. (1987, p. 109) concluded that ‘the origin of the Sahara as a continental desert . . . may be said to stem from the Miocene Alpine orogeny and the subsequent stripping of the Eocene deep weathering profile’. Sudano-Guinean woodland covered much of the Sahara during the Oligocene and early Miocene, having replaced the equatorial rainforest of Palaeocene and Eocene times. During the late Miocene and early Pliocene, a xeric flora, well-adapted to aridity, began to replace the earlier woodland, so many elements of the present Saharan flora were already present during the late Pliocene, when aridity became even more severe across the Sahara and the Horn of Africa (Bonnefille, 1976; Bonnefille, 1980; Maley, 1980; Maley, 1981; Bonnefille, 1983). The combination of a reduction in plant cover and a trend towards more erratic rainfall had a profound impact on the late Cenozoic rivers of the Sahara (Griffin, 2006). Big rivers capable of carving large valleys became seasonal or ephemeral. Integrated drainage systems became segmented and disorganised. Wind mobilised the sandy alluvium into active dune fields. Dunes formed barriers across river channels that were no longer competent enough to remove them. Dust storms left the desert top-soils depleted in clay, silt and organic matter. The Sahara was now a true wilderness, as the Arabic word implies.

The late Cenozoic desiccation which created the largest desert in the world was a result of a number of factors. Northward drift of the African plate ultimately helped disrupt the warm Tethys Sea to the north, with its abundant supply of moist maritime air.

Northern Africa moved away from wet equatorial latitudes into the dry subtropics. Growth of the great continental ice sheets and cooling of the oceans led to a decrease in precipitation and an increase in the strength of the Trade Winds. At the start of the Oligocene, there was a sharp drop in sea surface temperatures, which had eventual global repercussions. The late Cenozoic desiccation of the Sahara was not a continuous process but took place in stages. Evidence from marine cores collected off the west coast of the Sahara indicates that the terrestrial climate over the Sahara was relatively cold and dry at 24–20, 18–14, 13–9.5, 7.5–5.3, 3.2–1.9 Ma and from 0.73 Ma onwards (Sarnthein et al., 1982). During these intervals, the ITCZ displayed seasonal migrations similar to those of today, meridional winds were stronger and zonal winds were weaker. River loads were much reduced between latitudes 10° and 28° N, and dust flux into the Atlantic was increased. During the intervening stages (20–18, 14–13, 9.5–7.5, 5.3–3.2 and 1.9–0.73 Ma), the climate seems to have been less arid, with intervals of intense river discharge into the ocean. These climatic fluctuations were superimposed on the long-term desiccation caused by the northward drift of the African plate during the Neogene. Apart from tectonic uplift, what other factors were responsible for this dramatic change from a landscape of lowland equatorial rainforest to one of bare, rocky inselbergs and desert dunes?

Three additional influences seem to have contributed to the late Cenozoic desiccation of the Sahara. These were the uplift of the Tibetan Plateau, the build-up of continental ice in Antarctica and the Northern Hemisphere, and the cooling of the world's oceans. We have already considered the possible causes of these phenomena in [Chapter 3](#). Our concern here is with their effects on the Sahara.

Uplift of the vast Tibetan Plateau was an important factor contributing to the late Cenozoic desiccation of the Sahara and was associated with the intensification or even the onset of the easterly jet stream that today brings dry subsiding air to the deserts of Arabia and northern Africa, including the Horn of Africa (Flohn, 1980). A significant change in the late Miocene climate is evident in East Africa (Cerling et al., 1997) and Pakistan. Quade et al. (1989) identified a major change in the flora and fauna of the Potwar Plateau in the Siwalik foothills of Pakistan between 7.3 and 7.0 Ma, which may be related to Himalayan uplift and is consistent with the intensification or perhaps the inception of the Indian summer monsoon. This interval is coeval with one of the drier intervals identified by Sarnthein et al. (1982) for the west Sahara.

Although remote from the Sahara, the accumulation of continental ice in Antarctica also contributed to Saharan desiccation. There were mountain glaciers in Antarctica early in the Oligocene, and a large ice cap was well-established by at least 10 Ma (Shackleton and Kennett, 1975). Continental ice was slower to form in the Northern Hemisphere but was present in high northern latitudes by 3 Ma, and possibly by 5 Ma or even well before then, with a rapid increase in the rate of ice accumulation around 2.5–2.4 Ma (Shackleton and Opdyke, 1977; Shackleton et al., 1984). Closure of the Panama Isthmus around 3.2 Ma paved the way for rapid accumulation of continental

ice sheets in high northern latitudes during the late Pliocene (Schnitker, 1980; Loubere and Moss, 1986; Prentice and Denton, 1988). Oxygen isotope evidence from deep sea cores indicates that the onset of major Northern Hemisphere continental glaciations at 2.4 ± 0.1 Ma (Shackleton et al., 1984) also coincided with cooling in high southern latitudes (Kennett and Hodell, 1986).

The 2.5–2.3 Ma temperature drop is also evident in the south-eastern uplands of Ethiopia (Bonnefille, 1983) and the dry northern interior of China, with the beginning of widespread loess accumulation in the Loess Plateau of central China dated to 2.4 Ma (Heller and Liu, 1982). In the north-western Mediterranean region, the presence of a Mediterranean vegetation adapted to winter rains and summer droughts is already evident at 3.2 Ma, but it is not developed in its modern form until about 2.3 Ma (Suc, 1984). Magnetic susceptibility measurements of deep sea cores from the Arabian Sea and the eastern tropical Atlantic also reveal a change in the length of astronomically controlled climatic cycles at this time. Before 2.4 Ma, the dominant cycles are the 23-ka and 19-ka precession cycles, but after 2.4 Ma, the 41-ka obliquity cycle becomes dominant (Bloemendal and deMenocal, 1989).

As the two poles became progressively colder, high-latitude sea surface temperatures also declined. As a result, the temperature and pressure gradients between the equator and the poles increased. There was a corresponding increase in Trade Wind velocities, and therefore in the ability of these winds to mobilise and transport the alluvial sands of the Saharan depocentres and to fashion them into desert dunes (Chapter 8). Higher wind velocities were also a feature of glacial maxima during the Pleistocene and, as we saw in Chapter 9, were responsible for transporting Saharan desert dust far across the Atlantic (Parkin and Shackleton, 1973; Parkin, 1974; Williams, 1975; Sarnthein, 1978; Sarnthein et al., 1981).

The late Cenozoic cooling of the ocean surface would have helped reduce precipitation in the intertropical zone. About two-thirds of global precipitation now falls between latitudes 40°N and 40°S and depends on evaporation from the warm tropical seas (Galloway, 1965a). The reduction in evaporation from the tropical ocean that was associated with the global cooling resulting from high-latitude continental accumulation and enhanced cold bottom-water circulation would also lead to reduced rainfall across North Africa.

18.4 Uplift and erosion of the Afro-Arabian dome

During the Oligocene, there was a prolonged phase of slow crustal doming of the region centred on the northern Red Sea and southern Levant to form the Afro-Arabian dome (Bowen and Jux, 1987). This approximately 1,500 km wide elliptical dome extended across Ethiopia and Yemen, with its long axis aligned from south-south-west to north-north-east over a distance of about 3,000 km, and was located above the Afar plume (Avni et al., 2012). The uplift led to reactivation of pre-existing faults and

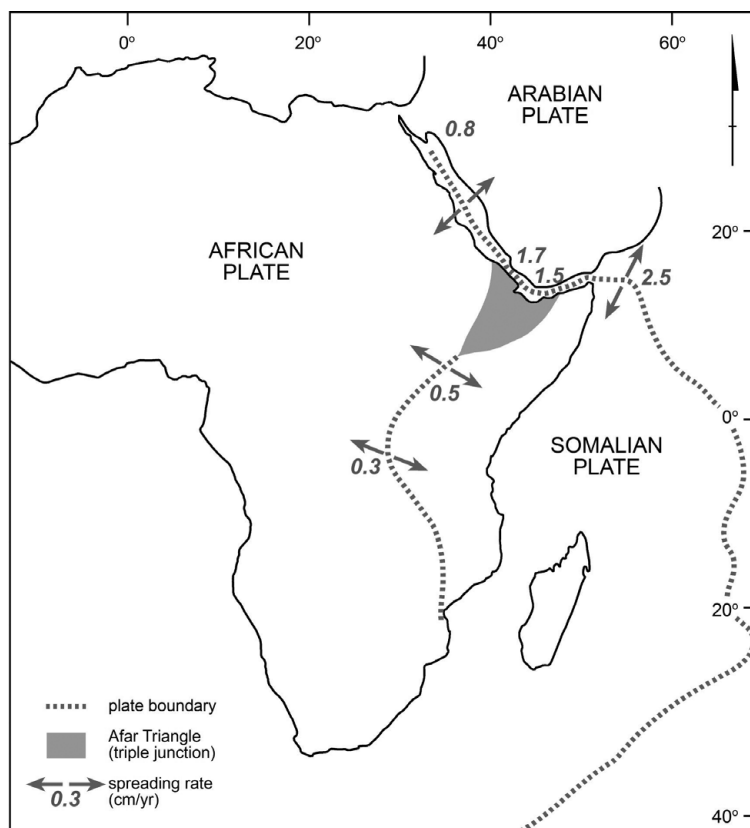


Figure 18.5. African and Arabian lithospheric plate movements and location of the East African Rift. (After Williams et al., 2004.)

tectonic lineaments, some of them already evident in the Precambrian basement rocks (Adamson et al., 1993). It also ushered in a long interval of erosion, estimated to have lasted for 6–10 million years, during which the Oligocene regional truncation surface was fashioned (Avni et al., 2012). Continued and accelerating upward expansion of the Afar plume during the late Oligocene and early Miocene caused rifting of the Red Sea, whose origin dates back to around 25 Ma. During the early–middle Miocene, the Dead Sea Transform originated as a left-lateral strike-slip plate boundary along a zone of pre-existing crustal weakness (Avni et al., 2012). Widening of the Red Sea Rift caused disruption of the initial Afro-Arabian plate and the formation of two separate plates: the African plate and the Arabian plate. The Red Sea is currently widening at a rate (from north to south) of 0.8–1.7 cm/year and the Gulf of Aden at 1.5–2.5 cm/year (Williams et al., 2004). Figure 18.5 shows current directions of movement of the African and Arabian plates.

The Afar Desert lies within the Afar rift and, together with Iceland, is one of the very rare places on earth where oceanic crust is forming today on land. The Afar,

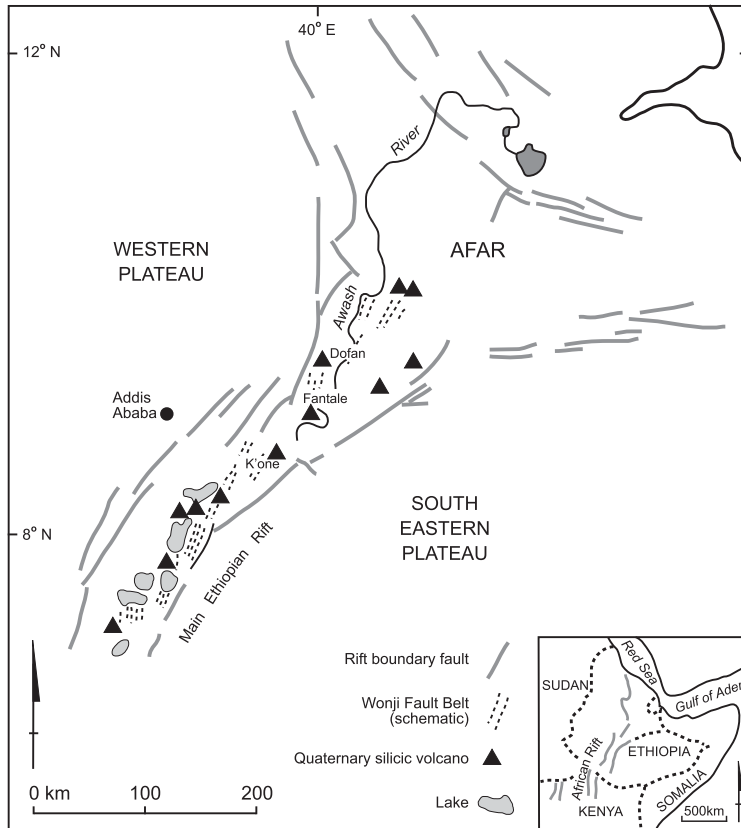


Figure 18.6. Main Ethiopian Rift, showing Quaternary lakes and volcanoes. (After Williams et al., 2004.)

sometimes termed the Afar Triple Junction because it is located at the intersection of the Red Sea, the Gulf of Aden and the Main Ethiopian Rift, is widening at a mean rate of about 0.5 cm/year and is also subsiding but at a much slower rate (Adamson et al., 1993). Miocene lake sediments and somewhat enigmatic Miocene granites occur within the Afar (Tiercelin, 1981; Adamson and Williams, 1987), but the most abundant rocks are basalts and occasional volcanoes, including the spectacular Erta' Ale volcano with its live lava lake. Pliocene and younger river and lake sediments contain a wealth of vertebrate fossils, including those of the hominids described in Chapter 17.

Rifting began in the Main Ethiopian Rift (MER) around 10 Ma ago, preceded by initial downwarping at around 15 Ma. Opening up of the 10–20 km wide Wonji Fault Belt, a zone of closely spaced, normal faults and fissures arranged *en échelon* within the 80 km wide MER (Figure 18.6), probably began about 1.6 Ma ago, and a major impulse of tectonic and volcanic activity has taken place within the last 0.25 Ma, after which the present MER lakes came into being. The current rate of widening of the MER determined from plate tectonic modelling amounts to about 0.5 cm/year, but

within the rift, currently monitored rates can be highly variable in time and space, ranging from as much as 0.1–0.45 cm/year to as little as 0.01 cm/year (Williams et al., 2004).

The three dominant directions followed by the Red Sea, the Gulf of Aden and the Main Ethiopian Rift are, respectively, south-east–north-west, west-south-west–north-north-east and south-west–north-east. All three are evident in the course pursued by the Nile River, with its sudden, sharp bends and long, linear reaches, and all three may be traced back to ancient lineaments within the Precambrian basement and overlying sedimentary rocks (Adamson and Williams, 1980; Adamson et al., 1992; Avni et al., 2012).

18.5 Cenozoic desiccation of East Africa

The late Cenozoic desiccation of East Africa, including Ethiopia, had several causes. Continuing tectonic uplift in East Africa during the past 6–8 Ma (Gani et al., 2007) created a major topographic barrier and caused a change in atmospheric circulation, reducing rainfall both in East Africa and the Chad Basin (Sepulchre et al., 2006). The change in rainfall regime over East Africa resulted in a change from tropical forest to open grassland and woodland and was associated with the proliferation of the Pliocene hominids unique to Africa, discussed in Chapter 17 (Cerling et al., 1997; Williams et al., 1998; Sepulchre et al., 2006).

Uplift and rifting in East Africa created the Neogene sedimentary basins, with their unrivalled record of Pliocene and Pleistocene hominid evolution. It is possible that the emergence in this region of the early Pliocene hominids may be linked to the Messinian Salinity Crisis (5.96 to 5.33 Ma: Cosentino et al., 2013), during which the Mediterranean Sea dried out, refilled and dried out repeatedly, resulting in the creation of a salt desert and the genetic isolation of Africa from Eurasia (van Zinderen Bakker, 1978; Williams et al., 1998). Both van Zinderen Bakker (1978) and Williams et al. (1998) believed that the Messinian Salinity Crisis was probably closely linked to the glacial evolution of Antarctica (Mercer, 1978), namely, expansion of the West Antarctic ice sheet and the concomitant sea level lowering in the very late Miocene. This remains a working hypothesis.

The late Miocene Nile responded to this change in base level by cutting a gorge more than 1,000 km long and up to 2 km deep at its northern end. This prompts us to ask when the Ethiopian tributaries of the main Nile River (i.e., the Blue Nile and Tekazze rivers) first originated, since we know that the White Nile and its parent Lake Victoria are relatively young features of the African landscape and probably no more than about 0.3 Ma in age (Talbot and Williams, 2009; Williams and Talbot, 2009). In order to answer this question, we need to consider the impact of certain tectonic and volcanic events on the Ethiopian drainage system, which is discussed in Section 18.6.

It is often hard to distinguish the precise causes of environmental change. The causes and consequences of the late Pliocene build-up of ice over North America

and ensuing tropical climatic desiccation are better constrained than the older glacial events described in [Chapter 3](#), but they still offer scope for differing interpretations. The same is true of when and why East Africa became arid.

For example, Cane and Molnar (2001) proposed that closure of the Indonesian seaway 3–4 Ma ago as a result of northward displacement of New Guinea in the early Pliocene would have triggered a change in the source of water flowing through Indonesia into the Indian Ocean from previously warm South Pacific waters to cooler North Pacific waters. The concomitant decrease in sea surface temperatures in the Indian Ocean could have reduced rainfall over East Africa. However, it seems unlikely that closure of the Indonesian seaway was the sole cause of late Pliocene desiccation in East Africa, since this region derives its moisture from both the South Atlantic and the Indian Ocean. It is equally possible that the late Pliocene increase in aridity evident in East Africa and Ethiopia 3–4 Ma ago (Feakins et al., 2005) may have arisen from the closure of the Panama Isthmus and the northward diversion of the warm equatorial water which until then had flowed westwards from the Atlantic into the Pacific Ocean. The presence of warm, moist air over the North Atlantic, coupled with a decrease in insolation that was linked to increased orbital eccentricity and a decrease in the tilt of the earth's axis (leading to cooler high-latitude northern summers and milder northern winters), was a prerequisite for widespread and persistent snow accumulation over North America (Williams et al., 1998).

The rapid accumulation of ice over North America at 3.5–2.5 Ma was accompanied by global cooling and intertropical aridity, revealed in the drying out of the large late Pliocene/early Pleistocene tropical lakes and rivers of the Sahara and East Africa. The emergence of stone tool-making at this time in East Africa may have been an adaptation by our ancestors to the increase in seasonality and the need to diversify their sources of food protein. As we have seen in [Chapter 9](#) this was also a time of widespread loess accumulation in central China (Heller and Liu, 1982) and of the first appearance of stony desert plains in central Australia (Fujioka et al., 2005). The region around the Mediterranean also developed its now characteristic dry summer, wet winter climatic regime. The net effects of these late Cenozoic environmental changes were an increase in the temperature gradients between high and low latitudes, a more seasonal rainfall regime, a reduction in forests and the replacement of woodlands by deserts in North Africa, Arabia and Australia, and the emergence in Africa some 2.5 Ma ago of upright-walking, stone tool-making ancestral humans.

18.6 Cenozoic uplift of the Ethiopian Highlands and Blue Nile incision

The history of the Nile is closely tied up with tectonic, volcanic and climatic events in its Ethiopian and Ugandan headwaters (Talbot and Williams, 2009; Williams and Talbot, 2009). Between Lake Tana in the Ethiopian Highlands and the modern Sudan border, the Blue Nile is entrenched into a plateau about 2.5 km in elevation and has cut



Figure 18.7. Dissected 30-million-year basalt flows near the headwaters of the Blue Nile, Semien Highlands, Ethiopia.

through Cenozoic basalts (Figure 18.7), Mesozoic and Palaeozoic sedimentary rocks and Precambrian basement rocks to form a spectacular canyon that is more than 350 km long, up to 20 km wide and up to 1.5 km in depth. McDougall et al. (1975) obtained the first potassium-argon ages ($27\text{--}23\text{ Ma}$) for the horizontal basalt flows through which the river had cut. They also estimated that $100,000 \pm 50,000\text{ km}^3$ of rock had been eroded from the gorge, which drained an area of around $275,000\text{ km}^2$. This amounts to a mean denudation rate of $15 \pm 7.5\text{ m}^3\text{ km}^{-2}\text{ yr}^{-1}$. This value is very low for a tectonically active region of high relief and is more consistent with rates from undisturbed forested tropical lowlands (McDougall et al., 1975). Modern erosion rates in the headwaters amount to at least $120\text{--}240\text{ m}^3\text{ km}^{-2}\text{ yr}^{-1}$ which is an order of magnitude faster than the mean geological rate. The volume of sediment stored in the Nile Delta and its much larger submerged cone in the eastern Mediterranean is $150,000 \pm 50,000\text{ km}^3$, which is very similar to the estimated volume of rock eroded from the Blue Nile gorge, allowing for changes in bulk density (Nile deltaic sediments: 1.5 g cm^{-3} ; bedrock eroded from Ethiopia: 2.8 g cm^{-3}), as well as losses in solution (estimated at 30 per cent). Relative to the volume of the Nile cone, the amount of alluvial sediment stored in the Nile flood-plain is trivial, amounting to $100\text{--}600\text{ km}^3$ along the main Nile, 800 km^3 in the Atbara Fan and $1,800\text{ km}^3$ in the Gezira Fan between the Blue and White Nile rivers in central Sudan.

The discrepancy between the modern erosion rates and the long-term geological rates suggests that there has been episodic uplift of the Ethiopian Plateau interspersed with periods of prolonged tectonic stability. Pik et al. (2003; 2008) used thermochronology to test models of Ethiopian landscape evolution. They obtained apatite helium ages showing partial resetting of pre-existing basement rock ages resulting from burial of the basement rocks beneath a thick mantle of Trap Series flood basalts around 30 Ma ago, and they concluded that erosion of the Blue Nile gorge began as early as 25–29 Ma ago, confirming the results of McDougall et al. (1975), with erosion along the scarps flanking the highlands starting after 11 Ma (Pik et al., 2003). The major volcanic/tectonic divides in Ethiopia date to 30–20 Ma and were formed before the rifting and break-up of the original Ethiopian volcanic plateau, which commenced after 20 Ma (Pik et al., 2008; Corti, 2009).

Gani et al. (2007) used a digital elevation model to reconstruct initial topography before erosion of the Blue Nile gorge and compiled potassium-argon ages for the volcanic rocks from present-day eroded volcanic remnants. They concluded that the Blue Nile and its tributaries had eroded at least 93,200 km³ of rock from the Ethiopian Plateau since 29 Ma. Uplift occurred in three phases (29–10, 10–6 and 6–0 Ma), with erosion rates accelerating at around 10 and around 6 Ma. The inferred rapid increase in erosion at 6 Ma (Gani et al., 2007) is the same age as that of renewed volcanism in the Afar Rift and renewed movement along the Dead Sea transform fault. Adamson and Williams (1987) speculated that the movement of the Dead Sea transform fault at this time may have been triggered by the repeated loading and unloading of the Mediterranean seabed during the Messinian Salinity Crisis of 6.2–5.3 Ma. The results of Gani et al. (2007) support the concept of episodic uplift and erosion put forward by McDougall et al. (1975) and raise the possibility that much of the sediment in the Nile cone may be of Pliocene age and younger.

We can therefore conclude that the elevated Ethiopian Plateau, which ultimately controls the hydrology of the Blue Nile and its related river systems, has been in existence since the late Oligocene. The Blue Nile and Atbara rivers have been ferrying sediment across the lowlands of Sudan and Egypt and into the eastern Mediterranean throughout the last several million years, and possibly at intervals over the last 30 million years. No other model seems able to account for the equivalence in volume of the Nile cone and the bedrock eroded from the Ethiopian headwaters of the Blue Nile and Atbara rivers, first estimated nearly forty years ago by McDougall et al. (1975). Nevertheless, there are persistent claims for alternative courses for the Nile either across to the Red Sea or into the Chad Basin. Such claims lack supporting sedimentary evidence and so remain speculative working hypotheses. Some of the proposed early courses of the Nile in Egypt show a complete reversal in flow direction to the south-west (Goudie, 1985). While the possibility of periodic disruptions to the Nile drainage network cannot be ruled out, the most compelling evidence seems to indicate that the long-term supply of water and sediment to the main Nile over the past 30 million years was primarily from the Ethiopian Plateau.

18.7 Quaternary environments in the Sahara and adjacent areas

The transition from somewhat wetter climates in the Miocene and Pliocene to the aridity that prevailed intermittently throughout the Quaternary in the African and Arabian deserts seems to have occurred in a series of stages. The African faunal record and the corresponding marine records for the last 5 million years reveal a series of wet and dry episodes linked to orbital variations (deMenocal, 2004). These fluctuations were superimposed on step-like changes to aridity and greater climatic variability at around 2.8, 1.7 and 1.0 Ma, coeval with the onset and intensification of high-latitude glacial cycles. The African faunal evidence is consistent with more open habitats at 2.9–2.4 Ma and after 1.8 Ma.

We begin with Jebel Marra volcano (Chapter 15, Figure 15.5), which lies at the geographical centre of the Sahara and is more than 1,500 km from the nearest coast in any direction. A large explosion caldera occupies the centre of the volcano. Two lakes occur today on the caldera floor. One lake is about 2.5 km long and was 11.6 m deep after the long wet spell of the 1950s and early 1960s but very shallow by January 1976. It is highly saline and alkaline. The other lake is about 1 km in diameter and up to 108.8 m deep. It occupies an explosion crater within the main caldera and is moderately saline (Hammerton, 1968). Jebel Marra and its surroundings provide a convenient case-study showing the different types of evidence that are used to reconstruct past environmental and climatic changes and some of the problems involved in dating and interpreting these events. Jebel Marra has a unique flora with a mixture of both southern tropical and northern Mediterranean species, indicating that it has been periodically connected with both margins of the Sahara (Wickens, 1976a). Miocene and Pliocene uplift has meant that erosion has long been active. As a result, the surrounding plains are mantled in Quaternary and older sediments. In addition, sporadic volcanic eruptions have continued into the Holocene. Reconnaissance mapping within a 100 km radius around the base of the volcanic massif revealed the following sequence of events (Williams et al., 1980; Philibert et al., 2010).

During the early Pleistocene, alluvial fans built up along the southern piedmont. They consisted initially of sands and gravels eroded from the Basement Complex rocks, which gave way to widespread water-lain tuffs following a phase of explosive volcanic activity and caldera formation. The tuffs contain fossil oil palm (*Elaeis guineensis*) and *Combretum* leaf impressions, and have Developed Oldowan/Early Acheulian artefacts in situ above and below them, with an estimated age between 1.5 and 0.8 Ma (Chapter 15, Figure 15.5). Extrusion of rhyolite lavas blocked valleys to the west and created a deep freshwater lake in which 5.5 m of laminated diatomite accumulated during a long wet interval (Philibert et al., 2010). Breaching of the lava dam caused the lake to drain. Upper Acheulian artefacts on the surface of the dried lake floor indicate an age of 1.0–0.5 Ma for this lake. Continued deposition of water-lain tuffs was followed by incision and formation of a 20 m terrace along

the southern piedmont. In southern Darfur, deposition of fluvial sands, mainly eroded from the Basement Complex but with minor volcanic inputs, provided sands that were reworked by wind to form the Older Qoz (dunes). Soils formed and stabilised the Older Qoz, which was breached by rivers during a wetter late Pleistocene phase. A delta formed within the caldera lake that was at least 25 m deep. The delta was later exposed and eroded and a wave-cut bench developed at 8–5 m above the caldera floor. Algal limestone deposits on the bench have calibrated ^{14}C ages of 22.5–19 ka. Maley (2000) obtained similar ^{14}C ages for lake and river sediments within Tibesti volcano and attributed them to changes in the Subtropical Jet Stream during the LGM. A sequence of alluvial terraces formed to the west and south of the massif at intervals during the late Pleistocene, with the presence of Middle and Late Stone Age artefacts indicating episodic fluvial deposition between 300 ka and around 15 ka. The Older Qoz sands were reworked by wind to form the (?) late Pleistocene Younger Qoz, which was then breached by rivers flowing from Jebel Marra piedmont. In southern Darfur, there was widespread deposition of a fining-upwards alluvial sequence, followed by the formation of a late Pleistocene 4 m silt terrace with LSA artefacts on the surface. During the Holocene, there was continued explosive volcanism and formation of a younger crater within the main caldera. Alternating silt deposition and river incision created the 3 m and 1.5 m terraces to the west and a 2 m terrace to the south of the massif. In southern Darfur, dark cracking clays (vertisols) were deposited during the early Holocene, when the climate was wetter than today. Incision ensued, followed by eolian sand and fluvial silt deposition. In the far south, large paleochannels were progressively filled with alluvium, and underfit channels were formed in the mid-Holocene. Incision below the modern channel beds was followed by deposition of coarse bed load sands within the channels along the western and southern piedmonts.

The most striking feature of the Jebel Marra alluvial record is the presence of oil palm fossils. The southern piedmont of Jebel Marra is today semi-arid, but the oil palm fossils show that tropical rainforest flourished there at some time between about 1.5 and 0.8 Ma. Another unusual feature is the former deep freshwater lake west of the massif, with its 5.5 m of finely laminated diatomites (Philibert et al., 2010). A thin layer of sandy alluvium with sporadic basalt and trachyte gravels overlies the diatomite. The artefacts on top of the exposed lake bed are typologically similar to the stone tool assemblages excavated at Olorgesailie in Kenya and from Olduvai Bed IV in Tanzania, which straddle the Brunhes-Matuyama paleomagnetic boundary (0.78 Ma) (see Chapter 6, Figure 6.2), which means that an age range of around 0.8 ± 0.3 Ma is likely for the stone tool assemblage (Isaac, 1982; Gowlett, 1984; Owen et al., 2008). Acheulian occupation sites elsewhere in northern and eastern Africa are always associated with the presence of freshwater lakes, rivers or springs (Clark, 1980).

There have been relatively few studies of older lake deposits in the Sahara, and they have often proved hard to date (Karim, 1968; Williams et al., 1980; Petit-Maire, 1982;

Williams, 1984; De Deckker and Williams, 1993; Wendorf et al., 1993). Gaven et al. (1981) and Petit-Maire (1982) considered Pleistocene Lake Shati in south-east Libya to be about 130 ka in age, that is, Marine Isotope Stage 5 (MIS 5). However, Williams (1984, p. 440) pointed out that the uranium-series (U-series) ages obtained from this site by Gaven (1982) on *Cerastoderma glaucum* shells showing little or no recrystallization (Icole, 1982) in fact fall into four distinct groups: 173–158, 136–132, around 90 and 40 ± 2 ka, suggesting that there were four lake phases and not just the single episode inferred by Petit-Maire (1982) and Gaven et al. (1982).

Dating lake carbonates is seldom a straightforward exercise. Causse et al. (1988) corrected for the effects of detrital thorium and obtained U-series ages of 100–80 ka for lake sediments in the west Sahara that were thought to belong to the last major wet phase in that area, regarded until then as early Holocene. Szabo et al. (1995) reported U-series ages for lake carbonates from Bir Sahara and Bir Tarfawi in the Western Desert of Egypt and other Pleistocene lakes in the eastern Sahara. They recognised five discrete lake phases dated to around 320–250, 240–190, 155–120, 90–65 and 10–5 ka. Crombie et al. (1997) obtained U-series ages on travertines from Kurkur Oasis in the Western Desert of Egypt that fell into three broad groups: >260, 220–191 and 160–70 ka. However, Wendorf et al. (1993, pp. 552–573) used a variety of dating methods to date a series of middle Pleistocene lakes at Bir Sahara and Bir Tarfawi associated with Acheulian and middle Palaeolithic artefacts, including luminescence (TL and OSL), uranium-series, amino acid racemisation and electron spin resonance. Only the OSL ages yielded stratigraphically consistent results, with ages between 175 and 80–70 ka for the various lakes and associated Middle Palaeolithic sites.

In the presently hyper-arid Murzuq Basin of southern Libya, at least four large lakes were present between 500 and 100 ka, with U-series ages of around 415 ka (MIS 11), 320–300 ka (probably MIS 9), 285–205 ka (MIS 7) and 138–128 Ka (MIS 5e) (Geyh and Thiedig, 2008). The lakes became progressively smaller during each successive wet phase, indicating progressively less humid interglacial conditions from MIS 11 onwards. Maxwell et al. (2010) identified other Middle to Late Pleistocene lakes immediately west of the Saharan Nile that were fed in part by overflow from the Nile.

In the Kenya Rift, Lake Naivasha showed three high lake level episodes between 175 and 60 ka (Bergner and Trauth, 2004). Before that time, during the late Pliocene and Early to Middle Pleistocene, Trauth et al. (2010) believed that the lakes of the Kenya Rift acted as amplifier lakes (see Chapter 11), showing an exaggerated response to minor changes in precessional forcing. The reason for this lies in the particular geomorphic setting of these lakes, which are located in low-lying sites with very high rates of evaporation but are fed from elevated catchments with high rates of precipitation.

Just as the occurrence of high lake levels may denote past intervals of higher rainfall, so too can the drying out of former lakes be used to reconstruct past episodes

of reduced rainfall. For instance, Cerling et al. (1977) used variations in the oxygen isotopic composition of pedogenic and groundwater carbonates (see [Chapter 15](#)) to infer a sharp reduction in precipitation at 2.0–1.8 Ma in the vicinity of Lake Turkana in northern Kenya and at 0.6–0.5 Ma around Olduvai Gorge in Tanzania.

As we approach the present, the evidence of former wet and dry episodes is increasingly well-preserved and more abundant. Remains of numerous late Pleistocene and Holocene lakes are scattered across the Sahara (Faure et al., 1963; Faure, 1966; Faure, 1969; Williams, 1971; Williams, 1973; Fontes et al., 1985; Ritchie et al., 1985; Pachur et al., 1990; Hoelzmann, 1993a; Hoelzmann, 1993b; Hoelzmann et al., 1998; Hoelzmann et al., 2000; Bonfils et al., 2001; Hoelzmann et al., 2001; Hoelzmann et al., 2004; Pachur and Altmann, 2006; Drake et al., 2011). The lake sediments often contain mollusc and ostracod shells, biogenic tufas and even charcoal (Williams et al., 1987; Gasse, 1990; Gasse, 2000a; Gasse, 2000b; Gasse, 2002), so dating these lakes by radiocarbon analysis is usually fairly reliable, provided that reservoir effects from ancient carbon can be assessed (see Chapters 6, 11 and 12). The majority of the younger former lakes in the central and southern Sahara are early to middle Holocene in age, with aridity setting in from around 5 to 4 ka and onwards. In the Afar Desert and the Kenya Rift, the late Quaternary lakes were low during the LGM, high for perhaps 10,000 years before then and high again during the early to mid-Holocene (Butzer et al., 1972; Gasse, 1975; Williams et al., 1977; Street and Grove, 1979; Williams et al., 1981; Gasse, 2000a; Gasse, 2000b; Chalié and Gasse, 2002). Zerboni (2005) analysed tufas in the Acacus massif, lake carbonates from the Edeyen dune field near Murzuq in the Fezzan and rock varnish from sites in the Messak massif. These localities are in the now hyper-arid Fezzan region of south-west Libya. The tufas showed that springs were active from 9.8 ka until the 8.2 ka cold event. The small lakes were highest between 10 and 8.2 ka and high again between 7.8 and 5.0 ka. Changes in rock varnish composition showed moist early to mid-Holocene environments, with dry conditions from 5.5 ka onwards and sustained eolian dust accretion in the past 1–2 ka (Zerboni, 2005; Zerboni, 2008).

More than thirty years ago, Rognon and Williams (1977) showed that events along the northern margins of the Sahara were not always synchronous with those in the central and southern Sahara and even in the southern Sahara there were regional variations linked to elevation. For example, from around 40 to 23 ka, lakes were generally high along the tropical southern margins, consistent with a wetter-than-present climate, glaciers were active in the Atlas and rivers flowed considerable distances from these mountains, in accord with a cool and wet full glacial climate along the temperate northern margins. From 23 to 15 ka, lake levels were in general low along the tropical margins, dunes were active and the climate was cold and dry. However, small lakes were present in Tibesti and Jebel Marra, perhaps fed by winter depressions. The summer monsoon became re-established after 14.5 ka and extended well into the central and eastern Sahara (Williams et al., 2006c), and lakes were

Table 18.1. *Late quaternary environments in the Sahara and Nile Basin (Modified from Williams, 2012a.)*

I. 250–25 ka

(Interval includes two glacial-interglacial cycles, each ~100 ka long)

- High White Nile floods at >240, 210 ± 30, 200 ± 40, 166 ± 30 and ~125 ka coeval with intervals of stronger Northern Hemisphere (NH) summer monsoon
- Sapropel accumulation in Eastern Mediterranean Sea at 217 ka (S8), 195 ka (S7), 172 ka (S6), 124 ka (S5), 102 ka (S4), 81 ka (S3) and 55 ka (S2), broadly coincident with phases of very high Nile discharge
- Lakes full and fresh during last two interglacial phases
- Integrated drainage across Sahara during last interglacial phase

II. 25–17 ka

(Interval includes the Last Glacial Maximum (LGM: 21 ± 2 ka))

- Lakes drying out in headwaters of Blue and White Nile rivers coeval with southward displacement of the ITCZ relative to today during NH summer and weakening of the summer monsoon; soils form on floor of Lake Albert at 20.7–17.7 ka and 16.5–15.1 ka
- White Nile deprived of overflow from Ugandan lakes and reduced to a seasonal trickle and its lower reaches blocked by desert dunes
- Lakes drying across the Sahara and East Africa
- Temperatures 4–8° C cooler than today in the Ethiopian Highlands, with local glaciation in the Semien and Bale Mountains and intense periglacial action and mass movement on mountain slopes; upper tree-line lowered ~1,000 m; bare unstable slopes supply abundant coarse debris to highly seasonal Blue Nile and Atbara rivers
- Widespread deposition of coarse sand and fine gravel across the Gezira alluvial fan in central Sudan and along the main Nile Valley in Sudan and Egypt; main Nile probably dries out during winter months; humans mostly abandon the main Nile Valley and migrate south in search of more reliable supplies of water
- Sand dunes active up to 500 km south of present southern limit of the Sahara
- Dust storms active, with export of desert dust to Europe, Negev Desert and Amazon Basin

III. 17–5 ka

- Lakes in Blue and White Nile source regions begin to rise at ~17 ka and overflow perennially at 15–14.5 ka
- Stronger summer monsoon and ITCZ extends ~500 km further north than today during the NH summer
- Upper catchments of Blue and White Nile densely vegetated and soil formation active
- Perennial channel flow re-established in Blue and White Nile and main Nile, which now carry a large seasonal suspension load of silt and clay
- Blue Nile incised >10 m into its former floodplain since 15 ka and >4 m since 9 ka, beheading its Gezira tributary channels, which dry out by ~5 ka
- High White Nile flood levels at ca.14.7–13.1, 9.7–9.0, 7.9–7.6, 6.3 and 3.2–2.8 ka; high Blue Nile flood levels at ca.13.9–13.2, 8.6, 7.7 and 6.3 ka
- Sahara is once more studded in sporadic lakes and supports a human population of Mesolithic hunter-fisher-gatherers and later Neolithic pastoralists
- A composite sapropel (S1) accumulates in the Eastern Mediterranean Sea between ~13.7 and 12.4 ka near the base and between ~9.9 and 8.9 ka near the top; the gap between the two sapropel subunits may denote the influence of the Younger Dryas (YD) episode

(~12.5–11.5 ka), which was marked by aridity in Lake Victoria and in the Sahara; flow in the Nile was probably curtailed and more seasonal during the YD

- Sudden decline in Saharan desert dust export to Atlantic at 15.5 ka, with low levels of dust flux until 5 ka

IV. 5–0 ka

- Lakes in Blue and White Nile source regions continue to overflow
 - Weaker summer monsoon and ITCZ retreats ~500 km to the south during the NH summer
 - More seasonal flow in Blue and White Nile rivers and main Nile; Blue Nile carries a mixed load of sand, silt and clay
 - Lakes dry out in the Sahara, which becomes abandoned by Neolithic pastoralists who move south into West Africa or east into the Nile Valley
 - Deposition of sapropel S1 in the Eastern Mediterranean Sea may have persisted until ~5 kyr, when the Nile deep-sea turbidite system also became inactive
 - Increase in Saharan dust flux to Atlantic
-

re-established, allowing Upper Palaeolithic hunter-gatherers and Neolithic herders to penetrate far into the Sahara. Along the northern margins, in contrast, the climate was mostly dry with minor wet intervals, rivers were less active and the plant cover indicated warmer-than-present conditions. After about 5 ka, the climate became drier along both margins and conditions became cooler. Long-term desertification set in, with human activities aggravating the ecological damage caused by more frequent droughts and sparser vegetation (Chapters 17 and 24).

In peninsular Arabia, the evidence from lakes is sparse (Rosenberg et al., 2011) and many of the so-called Holocene lakes may have actually been wetlands fed by groundwater with limited inputs from local run-off (Enzel, 2013). However, in contrast to the Sahara, there is a rich paleoclimatic archive in the form of speleothems from limestone caves in Oman and Yemen. As a general rule, speleothems will not form in desert caves unless there is a significant and reasonably prolonged increase in regional precipitation. Analysis of the oxygen isotopic composition and fluid inclusions preserved within speleothems dated by U-series has provided a 330,000-year record of wetter phases in northern Oman and southern Yemen (Fleitmann et al., 2003; Fleitmann et al., 2009; Fleitmann et al., 2011). Speleothem deposition was rapid at 330–300, 200–180, 130–120, 82–78 and 10.5–6.3 ka at Hoti Cave in northern Oman (Fleitmann et al., 2003; Fleitmann et al., 2009). Analysis of the speleothem D/H ratios (δD) and $\delta^{18}O$ values revealed that speleothem deposition coincided with interglacial or interstadial conditions during which groundwater was primarily recharged from moisture derived from the Indian Ocean when the monsoon rainfall belt extended further north and reached northern Oman. At Mukalla Cave in southern Yemen, speleothems likewise only formed during interglacial periods, such MIS 9, 7e, 7a, 5e, 5c, 5a and 1 (early to mid-Holocene) (Fleitmann et al., 2011). Maximum precipitation occurred during the last interglacial (MIS 5e) and the lowest precipitation occurred

in the early to mid-Holocene. Rosenberg et al. (2011) mapped four successive lake deposits in southern Arabia with ages of around 125 ka, around 100 ka, around 80 ka and early Holocene, comparable with the speleothem ages from Yemen and Oman. During any of these wetter intervals, prehistoric humans could have crossed the desert. There was a long dry interval in southern Arabia between around 75 ka and 10.5 ka, when human movement across the desert would have been very difficult.

The Quaternary speleothem records from the Negev Desert are more complex than those of Oman and Yemen and require more subtle interpretation. The $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ variations in speleothems from caves scattered in the southern and central Negev Desert have thrown light on past changes in surface vegetation and the probable source of the rainfall associated with speleothem formation. The speleothems indicate significant wet phases at 350–310, around 310–290, 220–190 and 140–110 ka, with all except the 310–290 ka humid phase coinciding with interglacial events (Vaks et al., 2007; Vaks et al., 2010). The wet phases were also coeval with episodes of sapropel accumulation in the eastern Mediterranean, discussed in Section 18.8 (see also Chapter 10), which occurred during times of greater discharge into the Mediterranean from the Nile and now defunct Saharan rivers. The 140–110 ka humid phase in the southern Negev was synchronous with the last interglacial wet phases evident in the speleothems in southern Yemen and northern Oman analysed by Fleitmann et al. (2003a; 2009; 2011). Nevertheless, only certain wet phases in the Negev Desert coincide with interglacial phases (e.g., 200–190, 137–123 and 84–77 ka), while others (e.g., 190–150, 76–25 and 23–13 ka) coincide with glacial phases (Vaks et al., 2006). The $\delta^{18}\text{O}$ values in speleothems from the northern Negev showed that the primary rainfall source was from the eastern Mediterranean, with some possible contribution from tropical southern sources during interglacial episodes. Vaks et al. (2003) found that during glacials, there was a southward migration of the desert boundary on the eastern flank of the central mountain ridge of Israel, but there was no change relative to the present during interglacials (Vaks et al., 2003). The D/H ratios in speleothem fluid inclusions showed that glacial climates were cooler with less evaporation over the eastern Mediterranean (Matthews et al., 2000; McGarry et al., 2004). Quantitative reconstructions of LGM temperature and rainfall are rare, but Affek et al. (2008) used ‘clumped isotope’ thermometry (see Chapter 7), and were able to show that temperatures in Soreq Cave south-west of Jerusalem were 6–7°C cooler than today during the LGM and 3°C cooler at 56 ka. One valuable by-product of such data is that they allow more rigorous testing of the climate models that are used in reconstructing global circulation during the LGM. The linear dunes in the southern Negev were more active during the LGM, when it was wetter than it is today, than they were during the more arid early Holocene (Roskin et al., 2011a; Roskin et al., 2011b). Although this may seem counter-intuitive, the explanation is simple: stronger winds and an abundant supply of sand at that time. This is in strong contrast to the Sahara and Arabia, where the dunes were mobile during the hyper-arid LGM and during the cold, dry

Younger Dryas ([Chapter 8](#)). Limited periglacial and glacial deposits in the Saharan and East African uplands are consistent with colder temperatures during the LGM in northern Africa, with estimated temperature lowering of 4–8°C relative to present (see [Chapter 13](#)).

18.8 Quaternary environments in the Nile Basin and adjacent areas

The Nile Basin occupies the north-east quadrant of Africa and contains a generous slice of that continent's climatic history. The Nile is the longest river in the world and carries a large volume of water and sediment from its tropical headwaters through Sudan and Egypt to the arid north-east coast of Africa. Since at least the Middle Pleistocene, it has acted as a corridor for and occasional barrier to human dispersal. Holocene floods fostered the advent of Neolithic farming in the Nile Valley and the subsequent emergence of one of the world's greatest urban civilizations. The Nile Basin also provides a unique offshore record of global climatic history. Marine sediment cores collected from the floor of the eastern Mediterranean show a repetitive sequence of alternating calcareous muds with a significant content of Saharan wind-blown dust and dark, organic-rich sediments termed sapropels (Larrasoña et al., 2003; Ducassou et al., 2008). Each sapropel unit spans up to about 10 ka in duration and accumulated during times of enhanced freshwater inflow from the Nile and now inactive Saharan rivers (Wehausen and Brumsack, 1998; Scrivner et al., 2004). Sapropel 1 (S1) is coeval with the early Holocene wet phase evident across the eastern Sahara (Kuper and Kröpelin, 2006). Wetter intervals in the main Nile Valley appear to coincide with S2 (55 ka), S3 (81 ka) and S4 (102 ka) but are still quite poorly dated (Williams et al., 2010b). S5 (around 124 ka: Lourens et al., 1996; Kroon et al., 1998) coincides with a time when the Western Desert of Egypt was a lake-studded savanna occupied by Middle Palaeolithic hunter-gatherers (Wendorf et al., 1993). Recently mapped lakes and drainage channels in southern Egypt point to periodic overflow from the Nile into these former lakes (Maxwell et al., 2010), prompting us to ask what was happening upstream at that time. It was also a time when an integrated drainage network connected the Chad Basin with the Mediterranean, allowing free movement of *Homo sapiens* across the Sahara (Osborne et al., 2008; Drake et al., 2011).

Herodotus (ca. 485–425 BC) surmised correctly that the black alluvial clays deposited each year by the Nile came from Ethiopia and were quite unlike the red desert soils of Syria and Libya, but he was puzzled by the cause of the Nile summer floods (Herodotus, 1960). In fact, three different rivers contribute to this flood. The Blue Nile and Atbara rivers flow from the Ethiopian Highlands and, until dams were built on both rivers, used to contribute 97 per cent of the annual sediment load and 90 per cent of the peak discharge in August but only 17 per cent of the June low-season discharge to the main Nile (Garzanti et al., 2006). The White Nile provides 83 per cent of the low flow but only 10 per cent of the peak discharge to the main Nile.

The modern confluence of the Blue and White Nile is at Khartoum, and the main Nile begins at this point. When the unregulated Blue Nile was in flood in modern times, the waters of the Blue Nile used to travel 300 km up the White Nile to create a 'flood reservoir' or 'pulsating lake' up to 3 km wide near its northern end (Willcocks, 1904). Once the Blue Nile flood had slackened, the pent-up waters of the White Nile were released into the main Nile, helping maintain perennial flow in that river. With the completion in 1935 of the Jebel Aulia Dam on the White Nile 35 km upstream of Khartoum, the reservoir when full also began to produce a body of slack water that extends about 300 km upstream. The creation of a seasonal lake in the lower White Nile Valley has most likely been the rule ever since the White Nile joined the Blue Nile during the Middle Pleistocene some 300,000 years ago (Talbot and Williams, 2009). However, the location of the confluence has shifted over time, as has the size and northern terminus of the lake. During the LGM at around 21 ka, aridity prevailed across intertropical Africa (Hoelzmann et al., 2004). Lake Victoria was dry, Lake Albert was low and there was no overflow from Uganda into the White Nile, which dried up (Adamson et al., 1980; Beuning et al., 1997). The abrupt return of the summer monsoon at 14.5 ka led to overflow into the White Nile and widespread flooding across the valley up to an elevation of 382 m relative to the Alexandria datum that was in use until recently (Williams et al., 1982; Talbot et al., 2000; Williams et al., 2006c). Lake Tana in Ethiopia also overflowed at that time (Lamb et al., 2007; Marshall et al., 2011), and the enhanced Blue Nile floods would have caused major flooding in the lower White Nile Valley (Williams et al., 1982). The flow of water into the White Nile was supplemented at that time by overflow from Lake Turkana into the Pibor and thence into the Sobat, a major White Nile tributary (Harvey and Grove, 1982).

There is striking evidence of even more extensive flooding in the White Nile Valley in the form of a lake that was more than 500 km long and up to 50 km wide at 386 m elevation (Williams et al., 2003). The 386 m strandline of this lake has recently been directly dated to last interglacial time using ^{10}Be cosmogenic nuclides (Barrows et al., 2014). With enhanced flow from both the Blue and the White Nile at that time, a seasonal lake formed and extended at least as far south as the present Melut bend on the White Nile. The Blue Nile floods would have propagated about 400 km upstream in the White Nile, to an elevation of 386 m. With the release of flow as the Blue Nile flood waned, sandy channel bars and dry-season sand dunes may have acted as temporary dams. The northern limit of the 386 m lake coincides with a former Blue Nile channel that bifurcates and joins the White Nile between 70 km and 120 km upstream of the modern confluence.

A series of former Blue Nile channels radiate north-west across the alluvial plain west of the Blue Nile towards the White Nile and vanish beneath the dunes located between Jebel Aulia in the north and Naima in the south (Chapter 10, Figure 10.10). The heavy mineral assemblage at six sites within the sand dune complex indicates that

the parent sediments originated in the volcanic highlands of Ethiopia, were ferried down by the Blue Nile and were later reworked as source-bordering dunes (Williams and Adamson, 1973; Williams, 2009b). The channels consist of a veneer of clay over sands and gravels. The channel sands become older with depth and have OSL ages between 70 ka and 115 ka, indicating a prolonged phase of fluvial sand entrainment and deposition. Within the error terms, the ^{10}Be age of 109 ± 8 ka for the strandline is statistically similar to the OSL age of 115 ± 10 ka for the fluvial Blue Nile channel sands and is consistent with backfilling of the White Nile Valley from Blue Nile floodwaters brought in by the former channel during the last interglacial (Barrows et al., 2014). The dimensions of the 386 m White Nile strandline suggest that the last interglacial Blue Nile peak floods were more extreme than the 14.5 ka floods that gave rise to the 382 m lake in the lower White Nile Valley. However, the last interglacial in North Africa was not uniformly wet but instead showed significant climatic variation (Rohling et al., 2002).

Independent evidence that the peak of the last interglacial was significantly wetter than the early Holocene comes from four quite separate sources: (A) The speleothem records from the Negev and peninsular Arabia provide unequivocal evidence that the last interglacial was much wetter than the early Holocene (Vaks et al., 2006; Vaks et al., 2007; Vaks et al., 2010; Fleitmann et al., 2011). (B) The record of last interglacial lakes from southern Libya and the eastern Sahara also shows that the climate was far wetter than it was during the early Holocene (Wendorf et al., 1993; Geyh and Thiedig, 2008; Maxwell et al., 2010). (C) The rivers that flowed across the central and eastern Sahara during the last interglacial never attained the same degree of integration during the Holocene (Osborne et al., 2008; Drake et al., 2011). (D) Finally, global sea level was 5.5–9 m higher during the last interglacial (MIS 5e, around 125 ka), consistent with significant ice melting in Greenland and the Antarctic (Dutton and Lambeck, 2012). Age calibration of the Red Sea last interglacial sea level record with the fine resolution Soreq Cave speleothem chronology indicates peak sea level at 132–126 ka (Rohling et al., 2008; Grant et al., 2012), somewhat older than but broadly consistent with the OSL age of 115 ± 10 ka for the Blue Nile paleochannel and the ^{10}Be age of 109 ± 8 ka for the 386 m White Nile strandline. Higher Nile discharge evident in the Mediterranean sapropel record (Lourens et al., 1996; Wehausen and Brumsack, 1998; Larrasoña et al., 2003; Scrivner et al., 2004) accords with a warmer global climate and stronger summer monsoon during MIS 5e. The 109 ka ^{10}Be age is the first direct age for the 386 m strandline and represents the maximum extent of this ‘reservoir lake’. As the lake gradually receded, sandy sediments were laid down between the high strandline and the present White Nile (Williams et al., 2003, fig. 2). Vertical incision in the main Nile after the last interglacial (Butzer, 1980, Williams et al., 2010b) would have lowered the White Nile base level from 386 to 382 m by the terminal Pleistocene. Renewed Nile incision in the mid-Holocene lowered the base level by a further 6–7 m on the main Nile (Bell, 1970) and 8–10 m at Khartoum

(Arkell, 1949). Incision by the Blue Nile amounts to at least 10 m since 15 ka, with associated incision by the White Nile of around 4 m since that time (Arkell, 1949; Arkell, 1953; Williams and Adamson, 1980; Williams et al., 2000).

The very gentle flood gradient of the White Nile (1: 100 000) has meant that the post-LGM flood deposits in the lower White Nile Valley are unusually well-preserved, in contrast to those of the Blue Nile. High White Nile flood levels have calibrated radiocarbon ages of around 14.7–13.1, 9.7–9.0, 7.9–7.6, 6.3 and 3.2–2.8 ka obtained on freshwater gastropod and amphibious *Pila* shells and fish bones (Williams, 2009b). The more fragmentary Blue Nile record shows very high flood levels at around 13.9–13.2, 8.6, 7.7 and 6.3 ka (Williams, 2009b).

Mayewski et al. (2004) examined fifty globally distributed paleoclimate records spanning the interval from 11.5 ka to present and identified six significant periods of rapid climate change at 9–8, 6–5, 4.2–3.8, 3.5–2.5, 1.2–1.0 and 0.6–0.15 ka. They observed that the first five of these episodes coincided with polar cooling and tropical aridity. The intervals in between were wetter in the tropics and are broadly similar to the intervals of high Blue and White Nile floods identified here. At the site of Erkowit in the Red Sea Hills (Mawson and Williams, 1984), there is evidence of permanent stream flow around 1.8–1.6 ka, coinciding with high White Nile flows but not as yet evident in the much more incomplete Blue Nile sedimentary record.

Verschuren et al. (2009) identified four episodes of low Holocene lake levels at Lake Challa on the slopes of Kilimanjaro with ages of 8.0–6.7, 5.9–4.7, 3.6–3.0 and 0.7–0.6 ka. These ages also roughly coincide with times of low flow in the White Nile (Williams, 2009b). The dry interval starting at 3.6 ka at Lake Challa may be coeval with the sharp decrease in rainfall along the southern Dead Sea at around 3.9 ka (Frumkin, 2009), although this may simply be a coincidence.

Unlike the sinuous suspension load channel characteristic of the early Holocene Blue Nile, the LGM Blue Nile and main Nile were highly seasonal rivers which carried a substantial bed load of sand and gravel, much of which they deposited in northern Sudan and southern Egypt during the long dry season, when they lost the competence to transport coarse sediment (Adamson et al., 1980; Williams, 2012a). A prime reason for this was the LGM desiccation of Lake Victoria and the curtailment of flow in the White Nile to a trickle. The White Nile today maintains perennial flow in the main Nile during years of drought in the Ethiopian headwaters of the Blue Nile and Atbara rivers.

18.9 Cenozoic evolution of the Namib and Kalahari deserts

The geomorphic evolution of southern Africa appears to have been far less complex than that of the Sahara, Afar and Negev deserts described earlier in this chapter. From west to east, the major elements of the landscape are the Namib plains, the Namaqua highlands, the Kalahari Basin, the Drakensberg escarpment and the Natal

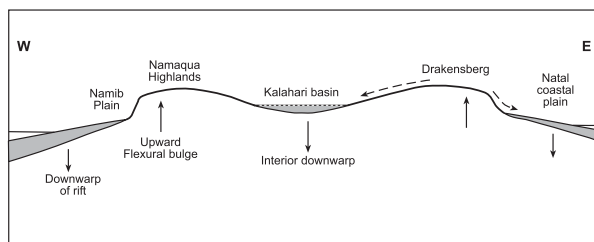


Figure 18.8. Generalised cross section across southern Africa. (After Thomas and Shaw, 1993.)

coastal plains (Thomas and Shaw, 1993) (Figure 18.8). The Kalahari Basin is an internally drained structural depression, with the major rivers flowing from the Angola highlands in the north-west down to the Okavango and Makgadikgadi Basin described in Chapter 11. The Kalahari sand sea covers about 2.5 million km² and is the largest sand sea on earth.

Measurements of ¹⁰Be and ²⁶Al indicate geological erosion rates in the hyper-arid Namib Desert ranging from about 3 to 5 m/Ma, consistent with minimal change in the Namib margin since the Eocene (Fujioka and Chappell, 2011). The Namib Desert lies between the Namaqua Highlands and the coast. This coastal desert is about 2,000 km long and 200–300 km wide. It owes its extreme aridity to three main factors. First, the presence offshore of the cold Benguela Current and associated cold upwelling coastal waters means that, apart from coastal fog, there is very little precipitation from westerly air masses, for the reasons outlined in Chapter 2. Second, the semi-permanent anticyclone located over the south-east Atlantic is associated with subsiding air and minimal convection. Third, the Namib Desert lies in the rain shadow of the Namaqua Highlands and therefore receives almost no rain from easterly sources.

As Siesser (1978) pointed out, the inception of extreme aridity in the Namib was determined by the onset of strong coastal upwelling. Sediment accumulation rates in marine cores off the coast of Namibia increased rapidly in the late Miocene, about 10 Ma ago, as did diatom productivity, isotopic evidence of colder temperatures and a sudden increase in phosphate formation (Siesser, 1978). All of these phenomena are consistent with enhanced upwelling, although there are some indications of earlier mild coastal upwelling starting in the late Oligocene. There is persuasive evidence that the progressive build-up of ice in Antarctica was the primary agent controlling the location and strength of the cold Benguela Current, with maximum ice accumulation in Antarctica at the end of the Miocene coinciding with a substantial increase in the upwelling of the Benguela Current (Siesser, 1978; Dingle et al., 1983; Coetzee, 1980).

Shackleton and Kennett (1975) were the first to identify a major drop in Southern Ocean temperature at the Eocene-Oligocene boundary based on isotopic analysis of

foraminifera collected from three marine sediment cores. This was also the time when a permanent ice cap was established in Antarctica (Chapter 3, Figure 3.4) and when glaciers first reached sea level around that continent (Mercer, 1978; Zachos et al., 2001). Dupont et al. (2005) obtained a continuous high-resolution pollen record from marine sediment cores in the south-east Atlantic for the interval between 3.5 and 1.7 Ma. They found evidence of rapid desiccation in the Namib at 2.2 Ma associated with a drop in sea surface temperatures and an increase in upwelling along the Namib coast. This contrasts with the marine records for the far south-west of Africa, where n-alkane $\delta^{13}\text{C}$ records indicate remarkable climatic stability over the past 2.5 Ma, preceded by a somewhat drier-than-present climate between 3.5 and 2.7 Ma (Maslin et al., 2012). N-alkanes occur in vascular plant leaf extracts and are quite resistant to degradation.

The northward movement of the African plate was also an important factor contributing to the onset of aridity in the Namib and Kalahari deserts, because it brought them into latitudes characterised by dry subsiding air and much-reduced precipitation (Habicht, 1979; Owen, 1983). Coetzee (1978; 1980) reviewed the pollen evidence for Cenozoic vegetation change along the south-west African coast. She noted that the late Miocene cooling in Antarctica coincided with an increase in upwelling of the Benguela Current and in the demise of palm vegetation. The 3.5 Ma drop in global temperature near the end of the Pliocene saw the disappearance of the last surviving remnants of the temperate forests, the spread of savannas and associated fauna, and the strong development of the winter rainfall *macchi* vegetation that is now characteristic of the south-west Cape.

18.10 Quaternary environments in the Namib and Kalahari deserts

Efforts to reconstruct Quaternary environments and associated climatic fluctuations in the Namib and Kalahari deserts using desert dunes have ranged from being exuberantly optimistic to cautiously pessimistic. With the advent of luminescence dating, it seemed to become possible to obtain ages for when dunes were actively forming and when they were stable. Stokes et al. (1997; 1998) used luminescence ages to infer multiple episodes of dune development (and presumed aridity) in southern Africa at 115–95, 46–40, 26–20 and post-20 ka. They interpreted the gaps in their age sequence as indicating more humid climatic phases during which the dunes became vegetated and stable. From this, they argued that changes in sea surface temperatures in the south-east Atlantic and Indian Oceans caused changes in temperature gradients and the movement of moist air masses into southern Africa from the north-east (Stokes et al., 1997; Stokes et al., 1998).

These pioneering endeavours were followed by a more rigorous approach to the luminescence dating of desert dunes, which soon demonstrated that apparent age clusters disappeared when the quartz sand samples were collected at closer vertical

intervals (Stone and Thomas, 2008). The greater abundance of OSL ages revealed that the dunes of the south-west Kalahari had been partly active throughout the last 120 ka and suggested that in this region, the dunes had been close to their threshold of reactivation throughout much of the late Quaternary, so earlier work invoking discrete phases of dune sand accumulation needed to be reassessed (Stone and Thomas, 2008). Furthermore, aridity may not have been the main cause of sand movement and could be outweighed by changes in wind velocity (Chase, 2009), which means that dunes are unreliable as indicators of past climate (Chase and Brewer, 2009). Perhaps what is now needed to resolve this impasse is a greater focus on the fossil soils within the dunes and any microfossils within them. Dunes will always be hard to interpret, because they reflect a number of different controlling agents, notably wind speed, plant cover, sand supply and effective precipitation, all of which could affect renewed dune movement.

Given the difficulties involved in using desert dunes to reconstruct past climatic changes, greater attention is now focussed on other lines of evidence, including lake deposits, speleothems and tufas, and pollen. The Makgadikgadi Lake complex at the distal end of the Okavango River in semi-arid Botswana has its headwaters in the equatorial highlands of Angola. This lake system is perhaps the best-dated set of lakes in semi-arid southern Africa, with more than 140 OSL ages now reported (Huntsman-Mapila et al., 2006; Burrough et al., 2009a; Burrough et al., 2009b). The Okavango high shorelines have OSL ages of 104, 92, 64, 39, 27, 17 and 8 ka, with small error terms of only a few ka. Older shorelines of 288, 267 and 131 ka have error terms between 25 and 16 ka (Burrough et al., 2009a). Hydrologic models suggest that above a certain size, the large lakes could have an influence on both local and regional climate (Burrough et al., 2009b). At present, when the Angolan highlands are wet, Botswana tends to be dry, and conversely. Huntsman-Mapila et al. (2006) considered that this anti-phase relationship between late Quaternary rainfall in southern and equatorial Africa also obtained during the late Quaternary. When Lake Ngami, which was fed by the Okavango, was high from 19 to 17 ka, there were signs of increased aridity in Botswana, which means that the LGM was arid in Botswana but wet in the Angolan headwaters of the Okavango.

Brook et al. (1997) compiled the ^{14}C and U-series ages of speleothems and tufas collected from the summer rainfall zone of southern Africa (Namibia, Botswana, northern Cape and the Transvaal) and compared them with the ages obtained for similar deposits across Somalia. Southern Africa was apparently wetter at 202–186, 50–43, 38–35, 31–29, 26–21 and 19–14 ka. Conditions there were wet during late glacial times and dry during the early Holocene. In Somalia, speleothem, tufa and rock-shelter sediments indicated wetter conditions in this presently arid region at 260–250, 176–160, 116–113, 87–75, 13, 10, 7.5 and 1.5 ka. From their survey, they concluded that over the last 35 ka at least, when it was wet in southern Africa it was dry in Somalia, and vice versa.

Scott and Woodborne (2007) used pollen preserved in hyrax dung to reconstruct late Quaternary vegetation changes in the winter rainfall area of south-west Africa. They deduced that the climate was cold and dry at 21–20 ka, warmer and wetter at 21–19 ka, cold and dry at 19 ka, wetter at 17.5 ka, warmer and drier at 16 ka, dry at 13–12 ka, warmer and wetter at 12–9.5 ka, slightly cooler at 11 ka and showing a decrease in summer rainfall at 5.6–4.9 ka. They noted that there could have been a slight increase in summer rain during the LGM and again possibly during the mid-Holocene, which would imply a possible southward displacement of the westerly wind belt, a conclusion contrary to that of Chase and Meadows (2007), which future work should resolve.

The marine sediment record has contributed greatly to our knowledge of late Quaternary (and older) climates in southern Africa along both the western and eastern coasts (Stuut et al., 2002; Stuut and Lamy, 2004; Stuut et al., 2004; Dupont et al., 2005; Pichevin et al., 2005; Schefuß et al., 2011; Maslin et al., 2012). Schefuß et al. (2011) analysed a marine sediment core located about 100 km off the Zambezi delta. They used the hydrogen isotopic composition of certain lipid biomarkers derived from higher plants to reconstruct the hydrologic changes in the Zambezi basin during the last 18,000 years. These changes closely reflected changes in $\delta^{18}\text{O}$ in the Greenland NGRIP ice core, showing that during cold events in the Northern Hemisphere such as the Younger Dryas (around 12.8–11.6 ka) and Heinrich stadial 1 (16.8–14.6 ka), rainfall was heavier and discharge was greater in the Zambezi drainage basin, probably as a result of the southward displacement of the ITCZ in the austral summer. Rainfall was also high in the last 4 ka, when local summer insolation was high. Pichevin et al. (2005) analysed the sediments in a 190 ka marine core off the southern Namib Desert and concluded that the mean accumulation rate of quartz grains could serve as a long-term indicator of aridity on land. They inferred that aridity in the Namib was greatest when summer insolation was low over the southern tropics. Stuut et al. (2002; 2004) and Stuut and Lamy (2004) also used changes in particle size in marine cores off the south-west coast of Africa to reconstruct past changes in Trade Wind strength related to changes in orbital precession, obliquity and eccentricity. An overriding problem with such an approach involves distinguishing eolian silts and fine sands from fluvial silts and fine sands. The issue is compounded by the presence in many Namib valleys of Pleistocene loess that has been reworked by running water and deposited as valley fills.

18.11 Conclusion

The long-term desiccation of the Sahara, Arabia and southern Africa is linked to the northward movement of the Afro-Arabian plate, uplift and volcanism in East Africa and southern Arabia, rifting, drainage disruption and the development of internally drained depressions. Oligocene and later uplift and erosion stripped away previous

deeply weathered mantles, creating a major erosion surface in north-east Africa and adjacent parts of peninsular Arabia. This surface was later disrupted by tectonic movement along zones of pre-existing crustal weakness, leading in stages to the formation of the Red Sea, the Gulf of Aden, the Afar Rift, the Ethiopian and Kenya rifts, and the Dead Sea Rift.

Superimposed on the trend of late Cenozoic desiccation, which is evident in the pollen record and that of large lakes and river systems, there were short-term climatic fluctuations. These were modulated by changes in orbital geometry, which controlled the length of glacial-interglacial cycles and changes in the strength of the summer monsoon. In intertropical Africa, glacial maxima were generally times of greater aridity and of temperatures 4–8°C lower than the present. In a number of instances, local hydrologic and topographic conditions counteracted the influence of regional changes in climate, as, for example, at Lake Masoko in Tanzania, which was relatively high when Lake Malawi nearby was relatively low during the LGM (Gasse et al., 2008).

In the southern Negev, desert dunes were active during the LGM, although the climate in that area was wetter than it is today. Stronger winds appear to have been the primary cause of the linear dune activity. In the Kalahari and Namib, the climatic signal conveyed by desert dunes is blurred by the influence of other factors, such as sand supply, wind strength and changes in surface cover, rather than aridity alone. Desert dunes were active along the southern margin of the Sahara during the cold, dry LGM and again during the cold, dry Younger Dryas. The LGM Nile was reduced to a seasonal trickle during the prolonged LGM drought, as were other big rivers, like the Niger and Senegal. The lakes in East Africa became shallow, and both Lake Victoria and Lake Tana in the White Nile and Blue Nile headwaters, respectively, dried out completely and soils formed on their exposed lake floors.

The abrupt return of the summer monsoon at 14.5 ka ushered in an era of plant and animal abundance, of widespread small lakes in East Africa, the Sahara and Arabia, and of Mesolithic hunter-fisher-gatherers and their cattle-herding Neolithic successors. Climatic desiccation set in anew from about 5 ka onwards, forcing the Neolithic pastoralists to migrate to wetter latitudes or favourable environments like the Nile Valley.

19

Asian deserts

Let us admit at once that we do not know what are the basic causes of climatic change.

Richard Foster Flint (1901–1976)
Glacial and Quaternary Geology
(1971, p. 789)

19.1 Introduction

This chapter draws together the disparate strands of evidence relating to climate change discussed in the earlier specialist chapters and seeks to provide a more integrated overview of the Cenozoic climatic history of the Asian deserts. The type of evidence on which this reconstruction is based includes desert dunes, desert dust and loess deposits, river and lake sediments and associated fossils, glacial deposits, soils, speleothems, marine sediments and stable isotope geochemistry. The late Quaternary environments receive most attention, because they are dated and documented in greater detail than earlier times. The record of historic floods and droughts is dealt with in [Chapter 23](#) and so will not be covered here.

19.2 Present-day climate and causes of aridity

[Figure 1.1 \(Chapter 1\)](#) shows the distribution of the arid, semi-arid and dry subhumid regions of Asia as defined and mapped in the UNEP *World Atlas of Desertification* (UNEP, 1997). To the north of these dry areas lie the cold temperate grasslands, woodlands and permafrost regions of Siberia. To the south and east, the woodlands and grasslands of the seasonally wet tropics give way to the rainforests of the perennially wet tropics. In contrast to the lowland deserts of Australia, Africa and Arabia, many of the Asian deserts are flanked by very high mountains, including many of the highest

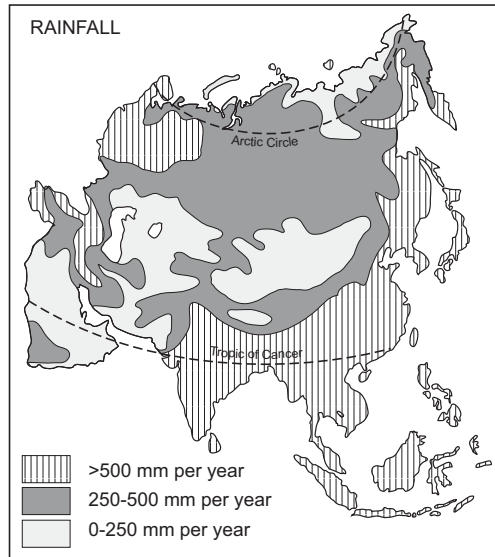


Figure 19.1. Mean annual precipitation in Asia. (After *The Times Atlas of Asia*, 2010.)

mountains on the planet, such as the Himalayas to the north of the deserts of India and Pakistan. As a result, many of the Asian deserts are located in the rain shadow of very high mountain ranges (e.g., [Chapter 8](#), [Figure 8.12](#)). In addition, the huge size of the continent dictates that rainfall in the interior is both sparse and erratic, because the moisture-bearing air masses from the ocean shed much of their precipitation within the first few hundred kilometres of the coast ([Figure 19.1](#)).

The climate of this vast region is dominated by the summer monsoon in the south and the winter monsoon in the north ([Figure 19.2](#)). Some precipitation is also derived from the mid-latitude westerly air masses from the Atlantic via the Mediterranean that bring modest amounts of snow and rain to the mountains north of the Tibetan Plateau, such as the Tian Shan. The strength of the Siberian anticyclone determines the strength of the winter monsoon. The summer monsoon is controlled by the pressure gradient between land and sea, which is in turn controlled by the temperature contrast between the ocean surface and the adjacent land. In fact, it is misleading to think in terms of a single, monolithic summer monsoon. In China, for example, the summer monsoon really consists of three quite independent monsoon systems, namely, the East Asian monsoon originating from the Pacific Ocean, the Indian monsoon originating from the Indian Ocean and the Plateau monsoon stemming from the Tibetan Plateau.

El Niño-Southern Oscillation events are an additional influence contributing to interannual precipitation variability in eastern China, northern Thailand and peninsular India, and they are discussed in detail in [Chapter 23](#). The sphere of influence of the

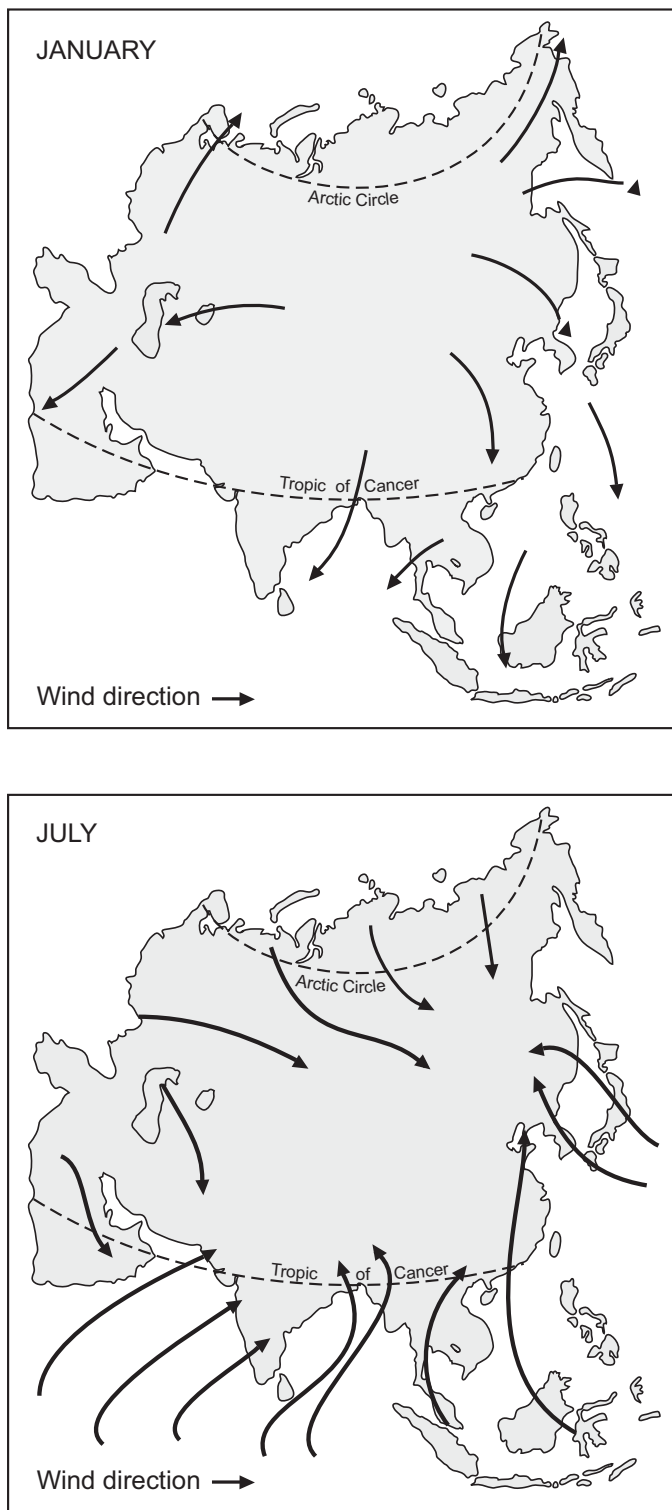


Figure 19.2. (a) Winter and (b) summer wind systems in Asia. (After *The Times Atlas of Asia*, 2010, and Kendrew, 1957, fig. 46.)

summer monsoon coincides very roughly with that affected by ENSO events ([Chapter 23](#); [Figure 23.1](#)). If a year of weak summer monsoon coincides with an El Niño year, the result will be a much lower than average summer rainfall. Conversely, if a La Niña year coincides with a year of strong summer monsoon, the result can be severe flooding as a result of unusually heavy summer rainfall, as in 2013.

Dust-storms are common in the drier parts of central Asia and China. Dust is mobilised during powerful convectional updrafts associated with strong frontal winds and is carried from eastern Siberia across eastern China, Korea and Japan out across the Pacific and sometimes as far as the Greenland ice cap, under the influence of high-level jet streams (Liu et al., 1985). Dust-storms originating in Mongolia and western China tend to occur most frequently during the passage of strong cold fronts in the Northern Hemisphere spring, when the Siberian high pressure system is weakening (Middleton, 1991; Roe, 2009).

19.3 Asian desert landscapes

All of the elements of desert landscapes outlined in [Chapters 1](#) and [2](#) are present in Asia but on the grandest of scales. The mountain ranges are huge, the desert dunes are the highest on the planet and the Loess Plateau of central China is the largest feature of its kind on earth. Cenozoic tectonic uplift and associated faulting were responsible for the creation of the Tibetan Plateau – the largest and highest plateau on earth, with a mean elevation of 4,600 m – as well as the Himalayas, Tian Shan, Kunlun Shan, the Pamir and Altai mountains, and many smaller ranges. Deep tectonic depressions between these ranges, such as the now almost waterless Tarim Basin bounded to the north by the Tian Shan and to the south by the Kunlun Shan, are occupied by hyper-arid deserts like the Taklamakan or the Badain Jaran Desert to the east, which is in turn bounded to the south by the Qilian Shan, a major supplier of sediment to the desert ([Chapter 8](#), [Figure 8.12](#)). In common with deserts elsewhere, the mountain ranges are flanked by alluvial fans but on a vast scale. These fans sometimes grade into gently undulating to nearly level elevated stony plains or plateaux, like the Alashan Plateau in north China's Inner Mongolia or the great Gobi Desert of northern China and Mongolia, which covers an area of 1.3 million km² and is the largest desert in Asia.

The sediments derived from the alluvial fans were reworked at intervals during the late Cenozoic to form the great sand seas of the Taklamakan, Badain Jaran and other deserts in China and, on a much smaller scale, the now mostly fixed sand dunes of the Thar Desert of India. Between these dunes, there are remnants of former river channels and occasional salt pans or small lakes. In the case of the Badain Jaran Desert, more than 100 lakes occupy hollows between the high dunes, and are fed by groundwater. Some of the salt lakes are huge, such as Lake Qinghai near the city of Xining in western China. The former pluvial Lake Lop Nor between the Gobi Desert to the

east and the Taklamakan Desert to the west was studied by the explorers Sven Hedin in 1899–1902 and Aurel Stein in 1906–1908, and both found evidence of wetter historic climates near the lake, which began to shrink soon after the third century (see [Chapter 5](#)). Berkey and Morris (1927) also reported evidence of previously more humid climates in presently arid Mongolia. In contrast to the evidence of once wetter climates in the form of now shrunken lakes and defunct river systems, the vast loess deposits of central China described by Baron Ferdinand von Richthofen (Richthofen, 1877–1885; Richthofen, 1882) seemed to argue for drier, windier conditions. The Silk Road from western China into central Asia and across to Europe was dependent on secure staging posts, of which the fabled oases of Bokhara, Tashkent and Samarkand are the best known. As in the Sahara and Arabia, these oases are situated in natural depressions watered from shallow groundwater, and so escape the tyranny of drought.

The drier parts of India display very different landscapes to those of central Asia and western China. The Thar Desert in north-west India and its southern margin in Gujarat contain a long record of past eolian, fluvial and pedogenic activity (Khadkikar et al., 2000; Chamyal et al., 2003; Juyal et al., 2006; Singhvi et al., 2010). The semi-arid sandstone plateaux of the Vindyan Hills and Kaimur Ranges that lie between the Ganga and Yamuna valleys to the north and the Son and Belan valleys to the south consist of sparsely wooded escarpments dissected by narrow valleys. The Son and Belan valleys contain an alluvial record extending back well-beyond the last interglacial 125 ka ago, as well as a record of human occupation from Lower Palaeolithic times onwards. Further to the west, the sandstone plateaux are capped by thick Cenozoic basalts, some of which are deeply weathered and capped by ferruginous duricrusts or laterites (see [Chapter 15](#)). Where such laterites occur in now dry areas, the former climate must have been much wetter.

19.4 Cenozoic tectonism, cooling and desiccation

Uplift of the Tibetan Plateau as a result of the collision of India and Asia around 45 Ma ago caused a major change in the distribution of land and sea and was followed by severe desiccation of the region to the north and east of the plateau, notably the Taklamakan, Badain Jaran and Gobi rain-shadow deserts ([Chapter 8](#), [Figure 8.12](#)). These deserts are still subject to earth movements and the deformation of Cenozoic and older sediments ([Figure 19.3](#)). The inception (or intensification) of the Asian winter and summer monsoons has also been attributed to these tectonic events, although other factors, such as the global cooling associated with the formation of permanent ice caps in Antarctica 34–33 Ma, may also have played a role. As Dettman et al. (2001) have pointed out, the Tibetan Plateau is a key driver of the Asian monsoon today. The surface of the plateau becomes hot in summer, the warm surface air rises and the ensuing low atmospheric pressure attracts moist air from the ocean, causing the heavy summer rainfall that is the hallmark of the Indian summer monsoon.



Figure 19.3. Cenozoic alluvial sediments tilted by recent tectonic activity, Xinjiang Province, north-west China.

In winter, opposite conditions apply, with cooling of the high-altitude plateau surface leading to high pressure and cold, dry winds blowing out across India, blocking any incoming moist air and minimising rainfall at that time (Figure 19.2). Because the Tibetan Plateau plays such an important role in generating the summer and winter monsoons, the history of the monsoon must be closely allied to that of the Tibetan Plateau, especially to when it attained sufficient elevation to have a major influence on the seasonal wind patterns.

Estimates for the inception of the monsoon extend as far back as the end of the Oligocene, with successive changes in monsoon activity postulated at 15–13, 9–8 and 3.6–2.6 Ma (An et al., 2001; Li et al., 2011). We saw in Chapter 18 that marine cores off the west coast of the Sahara show evidence of more humid conditions on land at 20–18, 14–13, 9.5–7.5 and 5.3–3.2 Ma (Sarnthein et al., 1982). There is a broad similarity between both sets of ages, suggesting that the summer monsoon influence may have extended as far as North Africa during those times.

The precise timing of uplift of the Tibetan Plateau is still a focus of research, and a variety of ingenious methods have been used in an attempt to clarify when uplift occurred in different parts of the plateau. Amano and Taira (1992) investigated the heavy minerals and rates of accumulation in early Miocene to Quaternary sediments in the Bay of Bengal deposited by rivers flowing from the Higher Himalayas. They

inferred two main uplift phases, one at 10.9–7.5 Ma and the other after 0.9 Ma. They also concluded that rivers in the upper Indus captured the headwaters of the Ganga between 7.5 and 6.5 Ma, probably as a result of tectonic movements. Burbank (1992) investigated Neogene sediments deposited by rivers coming from the Himalayas onto the Indo-Gangetic foreland at the foot of the Himalayas. He found a marked change in sediment accumulation in Pliocene-Pleistocene times, which was consistent with an increase in erosional unloading in the Himalayas during the past 4 Ma. Here we have a nice example of uplift triggered by denudation rather than by tectonic processes, showing that it is sometimes possible to distinguish between uplift arising from tectonic forces and uplift caused by erosion and, indirectly, by climate.

Li et al. (2011) analysed the neodymium (Nd) and $^{87}\text{Sr}/^{86}\text{Sr}$ isotope ratios in dust that was derived from Asia over the past 20 Ma and deposited downwind in the Pacific. They found from the change in isotope ratios that dust derived from the north Tibetan Plateau (NTP) showed an increase relative to that derived from the Central Asian Orogen after 15 Ma, and they concluded that the elevation of the NTP had increased gradually from about 2,700 to 4,500 m in the last 15 million years. The central and south Tibetan Plateau were already high by 15 Ma. Rohrmann et al. (2012) used a battery of thermochronologic techniques to establish that in central Tibet, the plateau had begun to form during the Late Cretaceous, expanding to cover much of central Tibet by 45 Ma. Methods used included apatite fission track dating and apatite [(U-Th)/He] dating of rates of exhumation and uplift.

Hetzel et al. (2011) used a combination of cosmogenic ^{10}Be exposure dating (see Chapter 6) and thermal modelling based on the (U-Th)/He ages of apatite and zircon to date a well-preserved peneplain situated at an elevation of about 5,300 m in the northern Lhasa block. They concluded that the peneplain had formed at low elevations until India's collision with Asia around 50 Ma ago resulted in crustal thickening, surface uplift and the subsequent preservation of the peneplain. Since Tapponnier et al. (2001) had already demonstrated that south Tibet had attained an elevation of at least 4,000 m by 35 Ma, Hetzel et al. (2011) inferred that there must have been rapid uplift of the Tibetan Plateau between the onset of the 50 Ma collision and 35 Ma.

Heller and Liu (1982) reported an age of about 2.4 Ma for the base of the wind-blown dust in the Loess Plateau of central China. However, this does not date the onset of aridity in this region. Sun et al. (2009) obtained an age of about 7 Ma for eolian dune sands in the central Taklamakan Desert, which is older than the age of 5.3 Ma for eolian siltstone on the windward edge of the Kunlun Shan flanking the southern Taklamakan (Sun and Liu, 2006). Further to the north-west in central Asia, the first evidence of eolian dust is far older, with strong evidence of aridity in that region by around 24 Ma (Sun et al., 2010).

Dupont-Nivet et al. (2007) obtained a fine-resolution magnetostratigraphic chronology for the Eocene-Oligocene transition (34–33 Ma) in the Xining Basin at the north-east edge of the Tibetan Plateau. Widespread sedimentation in playa lakes

persisted during the Eocene and ended abruptly at the Eocene-Oligocene transition, coincident with Cenozoic global cooling at 34–33 Ma, which was associated with the inception of permanent Antarctic ice sheets. The timing of uplift in Tibet is poorly constrained and probably time-transgressive, which prompted Dupont-Nivet et al., (2007) to conclude that the desiccation evident at the Eocene-Oligocene transition in this part of Asia is more likely to have been a result of global cooling than of regional tectonic events, although these events undoubtedly helped accentuate aridity. [A major drop in temperature is also evident 34–33 Ma in the Great Plains of North America (Zanazzi et al., 2007). We do not yet know with certainty when the Northern Hemisphere ice caps began to grow (Chapter 20). Drop-stones from ice-rafted debris laid down in the Norwegian-Greenland Sea between 38 and 30 Ma ago and apparently derived from East Greenland suggest that northern high-latitude ice accumulation may be far older than previously envisaged (Eldrett et al., 2007).]

The following year, Dupont-Nivet et al. (2008) presented the results of detailed pollen analysis at two sites in the Xining Basin that spanned the climatic transition from Eocene to Oligocene at 34 Ma. They noted the first appearance of the taxon *Picea* (*Piceapollenites*) in both sections, bracketed between 38.3 and 37.3 Ma, indicating a shift to a cool, temperate climate. They concluded that the appearance of coniferous taxa characteristic of high elevations at 38 Ma showed that there had been significant regional uplift of the Tibetan Plateau at least 4 million years before the Eocene-Oligocene transition, leading to enhanced silicate weathering and a concomitant decrease in atmospheric carbon dioxide (CO₂) concentration, as inferred by other workers (Zachos and Kump, 2005).

Late Cenozoic uplift elsewhere in the world, including the Rockies, the Andes and the Ethiopian Highlands, with concomitant erosion and weathering, would have accentuated the drawdown of atmospheric carbon dioxide. However, the sudden cooling at the Eocene-Oligocene transition 34 Ma ago is unlikely to be solely a result of the depletion in atmospheric CO₂. The abrupt temperature decline suggests a sudden change in boundary conditions, with the opening of Drake's Passage between South America and Antarctica being the most likely cause, because it enabled Antarctica to be girdled by the circum-Antarctic current without obstruction.

Another important factor also contributed to the long-term desiccation of Asia during the past 35 million years or so. The gradual shrinking during the Oligocene and Miocene of the warm and shallow Paratethys and Tethys seas that stretched across Eurasia was followed by a change from evenly distributed rainfall to a more seasonal rainfall regime. The Mediterranean is the last relic of the Tethys Sea and still contributes moisture to the arid lands to the east but not on the same scale as the former Tethys.

Quade et al. (1989) analysed the stable carbon isotopes preserved in fossil soils and fossil herbivore teeth in the Potwar Plateau of Pakistan (see Chapter 7). They found strong evidence of a major change in both flora and fauna between 7.3 and

7.0 Ma. Before 7.3 Ma, the dominant vegetation was forest and woodland. After 7.0 Ma, tropical grassland expanded rapidly as the forest cover dwindled. They suggested that these changes were consistent with the strengthening (or possibly even the start) of the Indian summer monsoon. The global increase in plants following the C_4 photosynthetic pathway (see [Chapter 7](#)) between about 8 and 6 Ma ago, and the corresponding decrease in C_3 plants are in accord with a decrease in the concentration of atmospheric CO_2 (Quade et al., 1989). C_4 grasses were already present in the Tugen Hills of northern Kenya by 15.3 Ma but did not become a major component of herbivore diet in Kenya and Pakistan until around 7 Ma ago (Morgan et al., 1994). The initial change from C_3 to C_4 grasses began first in the lowland tropics, because the threshold for C_3 photosynthesis is higher in warmer latitudes.

Dettman et al. (2001) built on the pioneering work of Quade et al. (1989) and analysed the stable oxygen isotopic composition of Neogene freshwater bivalve shells from Nepal, mammal teeth of similar age from Pakistan and soil carbonates from Nepal and Pakistan. The full record covered the last 11 Ma. They found evidence of a strong dry season signal from 10.7 Ma onwards, with little change in seasonal variability after 9.5 Ma. They also discovered evidence of significantly higher wet season rainfall before 7.5 Ma and drier conditions thereafter, which is consistent with the vegetation history and fossil soil evidence. They concluded that the Tibetan Plateau was already sufficiently high by at least 10.7 Ma to generate a strong summer monsoon.

19.5 Quaternary environmental fluctuations

19.5.1 Desert dust and loess

Very few terrestrial records span the entire duration of the Pleistocene. One outstanding exception is the Chinese loess record, which shows an alternation of loess accumulation in central China during cold, dry and windy climatic interludes and soil development (Kemp, 2001) under a re-established cover of moderately dense vegetation during the warmer, wetter intervals when the summer monsoon had become stronger once more. The dry intervals were coeval with glacial or stadial climatic phases, and the wet intervals corresponded to interglacial or interstadial phases (Liu, 1985; Liu, 1987; Kukla, 1987; Liu, 1991). The Loess Plateau of China (see [Chapter 9](#), [Section 9.6](#), [Figure 9.4](#)) occupies an area of about 440,000 km². The loess deposits have a mean thickness of about 100 m but locally attain 350 m and form some of the richest agricultural land in the world. The loess overlying the Pliocene eolian Red Clay in central China is Quaternary in age, but dust deposition has been active in central Asia from 24–22 Ma onwards (Guo et al., 2002; Sun et al., 2010), with significant dust accumulation in China by at least 8 Ma (An et al., 2001; Porter and An, 2005). The alternation of loess and well-developed soils with fossil pollen and mollusca indicative of woodland or forest (Kukla, 1987; Liu and Ding, 1998; Kohfeld

and Harrison, 2001) is consistent with alternating cold dry and warm wet climatic phases, and has been linked to the marine isotope record in the Pacific. Initial dating of the loess was based on the magnetic polarity time scale (see [Chapter 6](#)) supplemented by cross correlation based on magnetic susceptibility data (Kukla, 1987; Evans and Heller, 2001; Maher et al., 2010). Subsequent work has used radiocarbon and optical dating methods (Lu et al., 2007) for the upper part of the sequence, as well as correlation with the marine oxygen isotope record for at least the last 1 million years. Local dust sources inside China include the Taklamakan, Gobi, Badain Jaran, Tengger and Mu Us deserts ([Chapter 8, Figure 8.12](#); Ding et al., 1999; Pullen et al., 2011), with additional influxes from central Asia (Sun et al., 2010).

Traditional interpretation of the loess-paleosol succession invokes soil formation during times of stronger summer monsoon and loess accretion during times of stronger winter monsoon associated with a more intense Siberian high pressure system. Times of strong summer monsoon were equated with interglacial and interstadial phases, while times of strong winter monsoon were considered to be coeval with glacial and stadial episodes. Roe (2009) has questioned this interpretation on the grounds that present-day dust-storms in China occur mostly in spring, when the Siberian High is already weakening. In Mongolia, dust-storms are also mainly in spring (Middleton, 1991). However, soil formation requires a significantly wetter climate than that which prevailed during accumulation of the parent loess. Soils within the loess sequence are recognised on the basis of a wide variety of analyses, including grain size, magnetic susceptibility and micromorphology. To be ranked as a soil, they need to be at least as well-developed as the early Holocene soil at the top of the loess sequence.

Although the alternation between ‘winter’ and ‘summer’ monsoon may be somewhat oversimplified, it remains a useful model for future refinement. A more perplexing problem is that the loess sequence is not complete, and many sections from different sites still need to be studied. Porter and An (2005) drew attention to this issue after finding that interglacial phases often began with periods of severe gully erosion on the Loess Plateau. Loess is peculiarly susceptible to this form of erosion, so heroic efforts at hillside terracing are needed today to preserve the arable soils of this region.

Comparisons between the Chinese loess records, the oxygen isotope record preserved in marine sediment cores from the East China Sea, Pacific and North Atlantic, and the climatic record evident in Greenland ice cores (Liu et al., 1985; Kukla, 1987; Hovan et al., 1989; Porter and An, 2005) have been fruitful. They confirm existing climatic interpretations of the loess-soil couplets, with glacial maxima times of maximum dust deposition and interglacials with times of maximum chemical weathering of the loess and soil development. The loess deposits of Russia and central Asia show a similar sequence of alternating loess and soils, with loess accumulation during times of cold, dry, windy climate and widespread frost action (Rutter et al., 2003). The detailed studies of loess in China and elsewhere in Asia have demonstrated that

times of maximum dust flux were more arid than today and had a much reduced plant cover, an expanded dust source area and strong, very gusty winds (see [Chapter 9](#)). International recognition of the paleoclimatic importance of the unique Chinese loess archive stands as an enduring legacy of the pioneering efforts of the late Professor Liu Tungsheng.

19.5.2 Desert dunes of India and China

The Thar Desert of north-west India and eastern Pakistan ([Chapter 8](#), [Figure 8.10](#)) forms the eastern limit of the vast tropical deserts that extend from the Sahara across Arabia, Iraq, Iran and Afghanistan – a distance of 8,000 km spanning 110° of longitude, making this the largest stretch of deserts anywhere on earth. The Thar Desert itself covers an area of about 320,000 km². Mean annual rainfall diminishes from more than 500 mm in the east to less than 100 mm in the west, close to the Indus Valley. Here the linear dunes are aligned parallel to the dominant sand-transporting winds, which occur at the onset of the south-west summer monsoon. The desert is criss-crossed by ephemeral river channels indicative of once wetter times, when precipitation was enough to support perennial drainage. The seasonal Luni River is the only surviving integrated drainage system in the desert today.

In favourable circumstances, such as where the bedrock hills run at right angles to the dominant sand-moving winds, a long record of dune accumulation is preserved. For example, near Didwana in the north of the desert, one site shows twelve cycles of dune sand deposition, soil development, calcrete formation, erosion and renewed sand accumulation extending back to around 190 ka (Singhvi et al., 2010). Each full cycle lasted about 19,000 years, indicating a precessional influence. The onset of sand deposition coincided with the onset of monsoonal activity in this region, at least during the very late Pleistocene. Analysis of the carbon isotopic composition of organic matter within the sand profiles at this site indicated deposition of the sand during a transition from a landscape dominated by C₃ grassland to one covered in C₄ woodland or forest (Singhvi et al., 2010). This and other sites in the Thar Desert were occupied during more humid intervals from Lower Palaeolithic times onwards (Misra, 1983; Dhir et al., 1992; Dhir et al., 2010). In both the Thar Desert and the Wahiba Sands of Arabia, the period of most recent maximum dune building was not during the cold, dry and windy Last Glacial Maximum, as in the Rub al Khali and other nearby sand deserts ([Chapter 8](#), [Figure 8.9](#)), but at about the time when the south-west summer monsoon was again becoming stronger, around 15–14 ka (Wasson et al., 1983; Chawla et al., 1992; Dhir et al., 1992; Thomas et al., 1999; Singhvi et al., 2010). Dune activity in the Thar Desert and Wahiba Sands was therefore controlled more by wind strength than aridity.

Unlike India, which has only the one big desert located in an area of relative tectonic stability, China has twelve deserts or semi-deserts, many of them located in tectonic

depressions flanked by recently uplifted mountain ranges (Yang, 1991; Derbyshire and Goudie, 1997; Yang, 2002; Yang and Scuderi, 2010; Yang et al., 2011a; Yang et al., 2011b; Yang et al., 2012). Attempts to prevent or minimise dune encroachment on roads, railways and arable land led to the setting up of the Desert Research Institute in Lanzhou and to detailed mapping of dune type and rates of movement (Zhu et al., 1989; Zhu and Wang, 1992; Zhu et al., 1992; Zhu and Wang, 1993). In contrast to the tropical deserts of North Africa, Arabia and Australia, the Chinese deserts are located in mid-latitudes and so are subject to the mid-latitude westerlies. They are dry because they are located well inland. Flanked by high mountain ranges, they lie within zones of extreme rain shadow. The deserts west of the Helan Shan range lie within deep tectonic basins surrounded by very high mountains (see [Chapter 8, Figure 8.12](#)), many of which are capped by permanent snow and ice. Seasonal snow-melt feeds large rivers that flow into the deserts, where they vanish today but were more extensive in the past, especially during times when the dunes were inactive. Large alluvial fans at the foot of the mountains were and are a major sediment source for the active sand dunes. As a consequence, dunes cover a larger proportion of the more arid deserts in the west of China than is the case in North Africa, where high mountains are less common. For example, dunes cover 80–85 per cent of the arid Taklamakan in western China. This desert is bounded to the north by the snow-clad Tian Shan and to the south by the snow-clad Kunlun Shan, and is the largest desert in China, with an area of 337,600 km². Active dunes up to and slightly more than 100 m in height cover 80–85 per cent of its area. Very large lakes occupied the centre of the desert basin early in the Pleistocene, but they have since been buried or eroded, so little evidence is now left of these once wetter times (Yang et al., 2011b). At intervals in the late Pleistocene and mid-Holocene, rivers flowed through some of the dunes and fed small lakes, some of which persisted until about 300 years ago (Yang, 2001).

The second largest sand desert in China is the Badain Jaran (49,200 km²), bounded to the south by the ice-covered Qilian Shan and to the east by the Tengger Desert (42,700 km²) (Yang, 1991; Yang and Williams, 2003; Yang et al., 2010; Yang et al., 2011a; Yang et al., 2012). In the southern Badain Jaran Desert, the dunes are generally 200–300 m high. Some of the dunes are up to 460 m high and are the highest dunes on earth (Yang et al., 2011a). The reasons for this great height include the highly dissected bedrock beneath them, abundant fluvial sands from the Qilian Shan, a complex wind regime and the periodic stabilisation of the dune surface by calcareous soils during wetter climatic intervals. Between the dunes there are more than 100 small lakes, which are discussed in [Section 19.5.5](#).

Evidence from grain-size distribution, heavy mineral content and quartz grain isotope geochemistry has shown that individual deserts in China have been self-contained, receiving their sand supply from local rivers (Yang et al., 2012). As a consequence, they provide a local climatic signal that may not be representative of

the wider region. A great deal of work is still needed using lake sediments and fossil soils within the dune sands to determine past climatic fluctuations, while taking into account the often large reservoir effects (up to a few thousand years) that occur when using ^{14}C to date groundwater-fed lake deposits intercalated within the dunes (Hofmann and Geyh, 1998; Yang et al., 2011b).

19.5.3 River response to climate change

The big rivers of the drier parts of Asia contain a very partial record of Quaternary events and one that is often hard to decipher. For example, the headwaters of the Indus, Ganga, Yamuna and Huang Ho rivers have been subject to a variety of influences, including tectonic events, river piracy linked to such events, the waxing and waning of glacial activity and changes in monsoon intensity. As a result, it is difficult to derive a clear climatic signal from the often fragmentary alluvial record preserved in their middle and lower reaches (Srivastava et al., 2008; Srivastava et al., 2009; Ray and Srivastava, 2010). The sedimentary deposits of the Huang Ho in northern China also reflect fluctuations in the input and reworking of desert dust, so changes in rates of sedimentation may simply be a result of changes in dust inputs to the drainage basin. In the lower reaches of these big rivers, the influence of sea level fluctuations linked in particular to the growth and melting of the Northern Hemisphere ice caps adds further complexity to unravelling the alluvial history of these rivers. What is needed to obtain a climatic history is an alluvial record in valleys that are free from tectonic and sea level influences and have minimal inputs of wind-blown dust.

The river valleys of northern India contain a remarkably well-preserved record of alternating alluvial deposition and channel incision (Chamyal et al., 2002; Sridhar and Chamyal, 2010; Joshi et al., 2013). However, the relatively well-studied record of sedimentation in the Ganga (Srivastava et al., 2003) and Yamuna valleys reflects the influence of both climatic and tectonic factors in both of their upstream reaches and that of sea level fluctuations in the distal sector of the Ganga. South of the confluence of the Ganga and Yamuna rivers at Allahabad are two river basins that appear to be relatively free from both tectonic and sea level influences, namely the Son and Belan rivers (Williams and Royce, 1982; Williams and Clarke, 1984; Williams and Clarke, 1995; Pal et al., 2004; Williams et al., 2006b; Gibling et al., 2008). The alluvial deposits laid down by both rivers also contain a wealth of prehistoric stone tools ranging from Lower Palaeolithic to Neolithic, and these have helped to shed light on the transition from hunting and gathering to early agriculture in this region (Sharma, 1973; Sharma et al., 1980; Sharma and Clark, 1983; Clark and Williams, 1986; Clark and Williams, 1990). Both valleys contain four distinct alluvial formations, of which those in the middle Son Valley are the best exposed and best dated (Figure 19.4). From oldest to youngest, these are the Sihawal, Patpara, Baghor and Khetaunhi formations, with each type section named after a nearby village (Williams and Royce, 1982).



Figure 19.4. River terraces, middle Son Valley, north-central India.

The oldest formation in the Son Valley is informally known as the Sihawal formation, and is late Middle Pleistocene in age. It consists of a lower colluvial-alluvial member of locally derived quartzose sandstone blocks in a matrix of clay, and contains Lower Palaeolithic evolved Acheulian biface tools that are often very fresh. The lower member is overlain by a bed of silty clay that is entirely devoid of sandstone blocks and stone tools and may represent wind-blown dust. The lower unit appears to have been deposited by a combination of debris flows and alluvial fans at a time when the Son had entrenched its channel down to bedrock and had eroded any older alluvium from the valley floor (Williams et al., 2006b). The climate at that time was probably semi-arid with sparse vegetation along the valley sides and occasional strong downpours that were able to mobilise rocky debris and transport it downslope. The upper unit contains pollen of probable Himalayan origin, and if it is indeed a loess deposit, it would indicate a dry and windy climate, consistent with the absence of any prehistoric human presence in the valley at that time.

The Patpara formation is at least 10 m thick and, in addition to local quartz, sandstone and mudstone gravels, contains abundant pebbles of agate, chalcedony and other microcrystalline silicic rocks derived from the Deccan Trap basalts in the Son headwaters to the west. The matrix is a red clay. In places, the gravels are cemented by iron, indicating prolonged deep weathering under a humid tropical climate (see Chapter 15). A red-brown clay overlies the gravels, which contain a dominantly

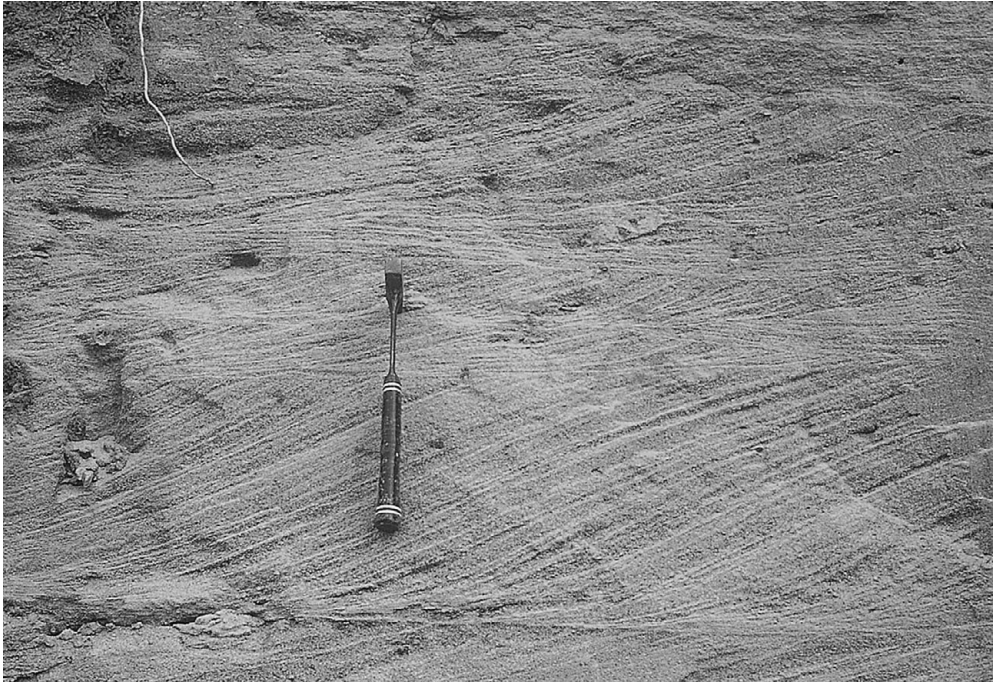


Figure 19.5. Cross-bedded late Pleistocene alluvial sands, middle Son Valley, north-central India.

Middle Palaeolithic stone tool assemblage. The age of this formation is early in the Upper Pleistocene, when the climate was far wetter than it is today, most likely during the last interglacial, when regional evidence points to a much stronger monsoon (Juyal et al., 2006; Gibling et al., 2008).

A pronounced erosional unconformity separates the Patpara formation from the overlying Baghor formation, which consists of a lower coarse member about 10 m thick and an upper fine member also about 10 m thick. The coarse member consists of cross-bedded medium to coarse sands (Figure 19.5) with lenses of sandstone, agate, chert and chalcedony ranging in size from granules to pebbles and discontinuous sheets of massive carbonate cemented sands. These latter beds contain well-preserved fossils of buffalo, hippo, antelope, elephant and tortoise, as well as rolled and abraded Middle Palaeolithic artefacts. The fine member rests conformably on the lower coarse member and consists of interbedded clays, silts and fine sands. It attains an elevation of at least 30 m above low water level and contains fresh Upper Palaeolithic artefacts in the upper few metres, as well as on the surface. This Upper Pleistocene formation is now reasonably well-dated using both ^{14}C and IRSL dating methods (Pal et al., 2004; Williams et al., 2006b). Deposition of the coarse member began at around 39 ± 9 ka and ended at around 16 ± 3 ka. This interval encompasses the cold, dry LGM, which was followed by a warmer, wetter climate after about 17–16 ka. The fine

upper member appears to have been deposited quite rapidly during a return to a stronger monsoon.

The Khetaunhi formation forms a low alluvial terrace up to 10 m thick and is banked against the older formations. It consists of interbedded silts, clays and fine sands and is overtopped during present-day floods. It contains Neolithic artefacts and was laid down between 5.5 ka and 3.5 ka.

There is some evidence in support of another formation informally termed the Khunteli formation (Williams et al., 2006b). This formation contains reworked volcanic ash from the 74 ka Toba super-eruption discussed in [Section 19.5.4](#). The ash overlies at least 6 m of medium sand, is up to 4 m thick and is overlain by up to 20 m of interbedded clays, sands and rolled carbonate gravels. The ash is a channel-fill deposit and has been locally eroded and replaced by several lenses of alluvial quartz and carbonate gravels. The absence of any deep weathering and iron precipitation in the alluvial sands and gravels suggests that it post-dates the Patpara formation. Carbonate cementation of the gravels above the ash points to drier conditions following the eruption, consistent with the $\delta^{13}\text{C}$ values in pedogenic carbonate nodules in fossil soils above the ash layer, which indicate that after the eruption grassland and open woodland replaced the former forest cover in this region (Williams et al., 2009a; Williams et al., 2010a).

While it is not possible to draw precise climatic inferences from the Quaternary alluvial deposits in the Son and Belan valleys, some provisional conclusions can be drawn. The phase of widespread and prolonged aggradation in the two valleys between around 39 ± 9 ka and around 16 ± 3 ka ended with sustained vertical incision and a return to warmer, wetter conditions. Renewed aggradation starting at around 5.5 ka marks a return to a weaker monsoon, more erratic rainfall and the onset of more frequent ENSO events (see [Chapter 23](#)) in this region (Williams et al., 2006b).

Enigmatic features of the landscape of the northern Thar Desert and Indus Basin are the abandoned cities of Harappa and Mohenjo-Daro. These cities formed part of the once flourishing Indus Valley Culture but were abandoned some 3,500 years ago. Located along the banks of now dry river systems, they have aroused debate as to why they were abandoned. Among the favoured hypotheses are invasion from the north-west, river capture, tectonic diversion of drainage and climate change (see [Chapter 12](#)). What seems reasonably well-established is that the Ghaggar-Hakra drainage system dried out after around 3.5 ka, at about the same time that the Thar Desert lakes were also drying out. Climatic desiccation therefore seems a likely cause of the abandonment of these cities.

19.5.4 Regional impact of the 74 ka Toba volcanic eruption

Few matters have been more widely debated in recent years than the possible impact on regional and global climate of the great explosive eruption of Toba volcano in

Sumatra 74,000 years ago (Williams et al., 2009b; Williams et al., 2010a; Williams, 2012b; Williams, 2012c). The precise age is 73.88 ± 0.32 ka (Storey et al., 2012). This eruption was one to two orders of magnitude larger than the historic and highly destructive eruptions of Tambora (1815), Krakatau (1883), Agung (1963) and Pinatubo (1991) (Chesner et al., 1991). Apart from killing many people, these historic eruptions were followed by a 0.5°C or so drop in mean global temperature, which persisted for one or more years (Rampino and Self, 1992; Kelly et al., 1996; Parker et al., 1996). Some recent eruptions have had additional effects, of which sustained cooling of the ocean surface is perhaps the most significant (Gleckler et al., 2006). The Pinatubo eruption in the Philippines was followed by a weakening of the regional hydrologic cycle and prolonged drought (Trenberth and Dai, 2007). A further apparent effect of the Pinatubo eruption was a change in the mode of the Arctic Oscillation (Stenchikov et al., 2002). (The Arctic Oscillation is controlled by surface atmospheric pressure and is reflected in periodic southward extensions of cold Arctic air into mid-latitudes, which have major impacts on North American weather).

In light of these impacts from quite minor volcanic eruptions, we might expect some significant impacts on regional and global temperature and precipitation arising from the 74 ka Toba eruption. Ninkovich et al. (1978a; 1978b; 1979) were the first to report ash from the 74 ka Toba eruption in marine sediment cores collected from the Bay of Bengal. Williams and Royce (1982) also found volcanic ash in the Son Valley of north-central India. This was the first time any Quaternary ash had ever been reported from India, and it soon led to discoveries of late Quaternary volcanic ash across the continent (Acharyya and Basu, 1993). Williams and Clarke (1995) compared the $^{87}\text{Sr}/^{86}\text{Sr}$ values preserved in the Son Valley ash with those obtained by Whitford (1975) from welded tuffs around the Toba caldera and found that they were identical and related to the most recent (74 ka) eruption.

Geochemical analysis of the volcanic ash by Shane et al. (1995; 1996) and Westgate et al. (1998) confirmed that the ash deposits across peninsular India all belonged to this 74 ka Toba eruption, which mantled India in a layer of ash 10–15 cm thick. In order to determine the possible impact of this event on the vegetation and climate of India and the wider region, the $\delta^{13}\text{C}$ values in pedogenic carbonate nodules in paleosols above and beneath the 74 ka ash layer were analysed across a 400 km transect in north-central India (Ambrose et al., 2007c; Williams et al., 2009b). Before the eruption, this part of India supported woodland and forest; after it, the forest was replaced by open woodland and grassland. Analysis of the pollen preserved in sediments above and beneath the 74 ka volcanic ash within a marine core from the Bay of Bengal revealed a similar pattern of vegetation change (Williams et al., 2009b). Other workers have concluded that the Toba eruption had little or no impact in India and the wider region (Gathorne-Hardy and Harcourt-Smith, 2003; Petraglia et al., 2007), a view that has not gone unchallenged (Williams et al., 2010a). However, as Williams (2012c) was at pains to point out, much of the existing work on Toba lacks the chronologic precision

to allow any robust inferences about its climatic impact to be drawn. Recent work in Greenland and Antarctica shows no evidence of sustained global cooling following the 74 ka Toba eruption (Svensson et al., 2013). However, on a more regional scale, the impact of the sudden deposition of a 10–15 cm layer of ash across peninsular India would have seriously disrupted photosynthesis and damaged the existing plant cover, leading to accelerated run-off and erosion, which is evident in the thick, channel-fill deposits of Toba ash in many parts of India. Whether or not it also caused human extinction in this region is for future work to determine.

19.5.5 Lake fluctuations

The great inland lakes of central Asia, such as the Caspian Sea, the Aral Sea and Lake Balkhash, all show evidence of having been much larger at some time during the Quaternary, and attracted the attention of scientific explorers more than 150 years ago (see Chapters 5 and 12). Detailed analysis of the sediments, microfossils (pollen, ostracods, diatoms) and stable isotopes associated with five lakes in northern Xinjiang and the arid Qinghai-Tibetan Plateau region show a sudden increase in summer rainfall at 12.5–11 ka that persisted until 8.7 ka, with the lowest levels in all lakes occurring between 4.5 and 3.5 ka (Fan et al., 1996; Gasse et al., 1996; Van Campo et al., 1996; Wei and Gasse, 1999). In the arid Badain Jaran Desert of Inner Mongolia in northern China, there are more than 100 permanent lakes in hollows between the dunes, often with well-preserved higher shorelines and fossil shells of freshwater mollusca. These shorelines have early to mid-Holocene radiocarbon and luminescence ages, and were previously much less saline than they are today, at a time when the local mean annual precipitation probably amounted to 200 mm, as compared to only 100 mm today (Yang and Williams, 2003). The lakes began to dry out after about 4 ka. Further to the north-east in the Ulan Bui Desert, a former lake reached its maximum extent between 7.8 ka and 7.1 ka (Zhao et al., 2012). The lake began to dry out after 6.5 ka. Until about 8.3 ka, this area was a sand desert; it now consists of a small salt lake (which supports a major chemical works) surrounded by active sand dunes.

The time of maximum Holocene rainfall varied across China, with several phases of strong summer rainfall during the early to mid-Holocene, followed by greater aridity after 5.5 ka, and particularly after 4 ka (An et al., 2000; Yang and Williams, 2003). The lack of synchronicity in Holocene lake fluctuations in China is partly due to the summer monsoon in China actually comprising three independent monsoon systems, as explained in Section 19.2. These are the East Asian monsoon coming from the Pacific, the Indian monsoon coming from the Indian Ocean and the Plateau monsoon coming from the Tibetan Plateau. In addition, certain of the deserts, such as the Badain Jaran, are in the path of the westerlies but are also close to the northern limit of the Asian summer monsoon along their southern margins.

The Pleistocene record of lake fluctuations in arid northern China and central Asia is less complete than the Holocene record. Some of the Badain Jaran Desert lakes were high around 34 ka, dry after 20 ka, and high once more by around 13 ka (Yang, 1991; Pachur et al., 1995; Yang, 2001b). In western Mongolia, the late Pleistocene lakes were higher than their Holocene counterparts, which were high at around 8.5 ka and around 1.5 ka (Lehmkuhl and Lang, 2001).

The Thar Desert of India has far fewer lakes than the Badain Jaran Desert of China. Singh et al. (1972; 1974; 1990) dated four lakes extending across a rainfall gradient from semi-arid east to arid west and analysed the pollen within the lake sediments. They found that after a long arid phase in the late Pleistocene, these presently saline lakes were fresh during the early to mid-Holocene and dried out soon after 4.5 ka, with the onset of desiccation starting a few centuries earlier in the arid western lakes than it did in the east. Singh (1971) suggested that the demise of the Indus Valley Culture was caused by this phase of desiccation, which also affected Mesopotamia, Persia and Afghanistan (Cullen et al., 2000; Weiss, 2000). Other workers have investigated several of the lakes first studied by Singh and his colleagues in greater detail. Wasson et al. (1984) carried out geochemical analyses of the sediments from Lake Didwana in the east of the Thar Desert, and Enzel et al. (1999) conducted detailed analyses of the sediments at Lake Lunkaransor in the arid west. Both studies showed that the early Holocene was wetter with a stronger summer monsoon. Overall, the early Holocene was a time of enhanced summer monsoon in the deserts of both India and China, while the Last Glacial Maximum was somewhat drier than today.

19.5.6 Glaciations

The use of ^{10}Be surface exposure dating of moraine boulders to date glacial advances in mountainous desert regions has revolutionised our knowledge of the glacial history of northern Mongolia, the Pamir and the Tian Shan (Zech et al., 2005; Gillespie et al., 2008; Sanhueza-Pino et al., 2011; Zech, 2012). We noted earlier that the Tian Shan forms the northern boundary of the Taklamakan Desert. These mountains are 1,500 km from west to east and rise to above 7,000 m. Westerly air masses from the Atlantic and Mediterranean bring more than 1,000 mm of precipitation to the northern and western slopes of the Tian Shan during spring and summer, but in winter the Siberian High blocks the flow of moist westerly air. The southern and interior slopes of the Tian Shan are arid, receiving less than 300 mm of rain a year, mainly from convective storms in summer. The accumulation of snow and ice in the Tian Shan is therefore limited by temperature in the north and west and by precipitation in the interior and south. The Pamir Mountains lie south of the Tian Shan and form the western boundary of the Taklamakan Desert. They receive some monsoonal rain from the south during summer. As a result of the different precipitation sources in

both mountain ranges, the glacial history of the Pamir differs from that of the Tian Shan. Zech et al. (2005) compared ^{10}Be surface exposure ages for moraines from the Tian Shan with those from the Pamir. The Tian Shan moraines had ages of 15, 21 and >56 ka (MIS 3), while the Pamir showed extensive moraines during the last interglacial (MIS 5) (Zech, 2012). The MIS 3 glaciers in the Tian Shan may indicate an increase in westerly precipitation at that time, with the more southerly Pamir receiving increased monsoonal precipitation during MIS 5. Ice was more extensive in both the Pamirs and the Tian Shan during MIS 4 than during MIS 2, suggesting an increase in aridity in Central Asia during the last glacial cycle (Zech et al., 2005; Zech, 2012).

Sanhueza-Pino et al. (2011) obtained ^{10}Be exposure ages of 67–63, 15–11 and 8–6 ka for boulders from very large landslides in three previously glaciated valleys in the Kyrgyz Tian Shan. These ages provide a minimum age for glacier advances in their respective valleys and also mark the maximum extent of the ice within those valleys up to the time of the landslides. The MIS 4 age for one of the glacial advances shows that the glacial history of these mountains cannot be widely extrapolated and also revealed that the minimum elevations of the equilibrium snow-lines (see Chapter 13) in the northern Tian Shan was about 400 m higher than previously estimated.

The two most recent glacial advances in the Darhad Basin of northern Mongolia had ages of around 53–35 ka (MIS 3) and around 19–17 ka (MIS 2) (Gillespie et al., 2008). In order to establish a reliable chronology of times of maximum local ice advance and times of ice retreat, three independent dating methods were used: ^{14}C , ^{10}Be and luminescence (IRSL) (see Chapter 6). The ages of these glacial advances were the same as advances of similar extent across northern Mongolia, but they differed from those of glacial advances in Siberia and western Central Asia. An older and more extensive glaciation may date to MIS 6, but there was very little difference in the equilibrium line altitude (ELA) for all three glaciations. Unlike the Tian Shan, where greater LGM aridity confined glaciers to high elevations, glaciers advanced to relatively low elevations during the LGM in the Darhad Basin of northern Mongolia, indicating that it was less arid at that time (Gillespie et al., 2008).

The Tibetan Plateau is the largest high plateau on earth, with an area of 2.6 million km^2 and a mean elevation of 4,600 m. Kuhle (2001; 2002) has argued that most of the plateau was covered in ice during the LGM, but the majority of workers have concluded that glaciation was limited and spatially isolated (Owen, 2009), not least because both pollen and $\delta^{18}\text{O}$ proxy data indicate that the Alpine Steppes of the Tibetan Highlands persisted during the LGM, when temperatures were about 3–4°C lower than they are today (Miehe et al., 2011). The Tibetan Plateau may have been slightly more arid than today during the LGM, but it was not excessively dry, or else the Alpine Steppes could not have survived.

19.5.7 *Speleothems*

Although the loess record from central China spans the entire Quaternary and provides the longest continental record on earth of past climatic changes in arid areas, its chronology is relatively imprecise. The speleothem records from various caves in China provide an invaluable adjunct to the more recent loess record and have the advantage of being very precisely dated. The caves are often located well outside the arid zone, but they still yield important information relevant to the deserts, especially in regard to past changes in the intensity of the summer and winter monsoons, which influence the desert margins. In addition, because they are very precisely dated using uranium-series dating methods (see [Chapter 6](#)), the speleothem records can be compared with high-resolution records in other parts of the world, including the Greenland and Antarctic ice core records, north-east Brazil, the Cariaco Basin off Venezuela, peninsular Arabia and the North Atlantic. Such comparisons help provide a more integrated view of global climatic fluctuations and possible forcing factors.

Wang et al. (2001) studied the $\delta^{18}\text{O}$ records from five stalagmites from Hulu Cave near Nanjing covering the past 75 ka. They found that the East Asian monsoon was more intense when Greenland temperatures were warmer and weaker during cold intervals in Greenland. Cheng et al. (2006) extended the Hulu Cave record back to the penultimate glacial and deglacial phases and discovered that both of the glacial terminations comprised two phases, with an interval of weak monsoon followed by an abrupt increase in monsoon strength. They also identified at least sixteen millennial scale events during the penultimate glacial period, comparable to the Dansgaard-Oeschger cycles of the last glacial period (see [Chapter 3](#)). Wang et al. (2008) later obtained a 224 ka $\delta^{18}\text{O}$ record based on 127 ^{230}Th ages obtained from twelve stalagmites in Sanbao Cave in central China. They deduced that changes in the strength of the East Asian monsoon reflect orbitally controlled variations in high northern latitude insolation, with a 23 ka periodicity punctuated by millennial scale events. Yuan et al. (2004) analysed the $\delta^{18}\text{O}$ record of precipitation changes over the past 160 ka from two stalagmites in Dongge Cave, 1,200 km west-south-west of Hulu Cave. These were consistent with those inferred from Hulu Cave and showed that the Last Interglacial Monsoon began quite abruptly at 129.3 ± 0.9 ka and ended equally abruptly at 119.6 ± 0.6 ka, consistent with the changes in temperature in the North Atlantic region that are evident in Greenland ice cores.

Speleothems have also been used to resolve the question of whether the East Asian summer monsoon (EASM) and the Indian summer monsoon (ISM) were in or out of phase. Winds blowing from the south-east flowing across the western Pacific and the South China Sea into eastern China and central China control the EASM. Winds blowing from the south-west across the Indian Ocean and Bay of Bengal into southern China control the ISM. Cai et al. (2006) obtained a high-resolution $\delta^{18}\text{O}$ record from a stalagmite dated between 53 and 36 ka from Xiaobailong Cave in south-east China

which showed that millennial scale variations in the ISM revealed in this cave were indeed synchronous (at least, at that time) with those from Hulu Cave in the path of the EASM.

One of the few cave records to come from the drier parts of China is an especially interesting $\delta^{18}\text{O}$ record from Wanxiang Cave in semi-arid north-west China, situated between the Qinghai-Tibetan Plateau to the south-west and the Chinese Loess Plateau to the east. The record covers the last 1,810 years (Zhang et al., 2009). The monsoon in this locality was strong during Europe's Medieval Warm Period and weak during the Little Ice Age. It was also weak during the final decades of the Tang, Yuan and Ming dynasties, when poor harvests triggered widespread unrest. Weaker monsoons coincided with times of decreased solar intensity.

An as yet unresolved question relating to interpretation of the speleothem records in terms of monsoon intensity concerns the precise nature of the climatic signal. Maher and Thompson (2012) have suggested that the $\delta^{18}\text{O}$ records from Chinese stalagmites may reflect changes in the source of the moisture rather than changes in the actual amount of summer monsoonal rainfall. However, changes in air mass precipitation sources will, in any case, cause changes in the amount of rainfall, so this may not prove to be a major problem.

19.5.8 Marine cores

The evidence from rivers, lakes and dunes is consistent with a drier climate during the LGM in India and further afield. One way to test this inference is to examine the isotopic composition of deep-sea cores. At a global level, changes in the $^{16}\text{O}/^{18}\text{O}$ ratio in ocean water reflect changes in the global ice volume (see Chapter 3), but at a regional level, changes in the $^{16}\text{O}/^{18}\text{O}$ ratio within calcareous marine fossils can be used to assess past changes in sea surface salinity and temperature.

Duplessy (1982) used this approach to estimate changes in salinity in the Bay of Bengal. He found an increase in salinity in the northern Bay of Bengal during the LGM, which he interpreted as resulting from a reduction in freshwater discharge into the Bay of Bengal because of less rainfall and more arid conditions at that time. In the Son Valley of north-central India, the presence of carbonate-cemented alluvium dated to 25–15 ka is also consistent with drier late glacial climates in this region (Williams and Clarke, 1984; Williams and Clarke, 1995).

Zhang et al. (2009) used changes in the hematite to goethite ratio in a deep-sea sediment core in the South China Sea to determine changes in monsoon intensity during the last 5 million years. The sediments were derived from the lower Mekong River. A high proportion of hematite indicates relatively dry conditions in the basin with minimal chemical weathering, while a high proportion of goethite denotes intense weathering under humid conditions. They found that during the early Pliocene, from 5.0 to 4.2 Ma, the monsoon was weak, becoming intense during the mid-Pliocene,

from 4.2 to 2.7 Ma ago, after which it became weaker again and more variable. They also found high concentrations of ^{44}Ca in the core sediments during the interval of intense monsoon, and concluded that high rates of chemical weathering and rapid denudation at that time caused a reduction in the atmospheric concentration of carbon dioxide (pCO_2) and triggered the onset of late Pliocene major Northern Hemisphere glaciation around 2.7 Ma ago.

Tian et al. (2002) offered a somewhat different interpretation of monsoon strength based on their analysis of the $\delta^{18}\text{O}$ composition of benthic foraminifera from marine sediments collected from ODP Site 1143 in the South China Sea. The core spanned the last 5 Ma. They concluded that the strengthening of the East Asian monsoon took place after 2.5 Ma, probably as a consequence of the increase in continental ice volume in the Northern Hemisphere.

One way to reconcile the apparently conflicting conclusions of Tian et al. (2002) and Zhang et al. (2009) is to propose that mid-Pliocene accelerated weathering and erosion in the headwaters of the large tropical Asian rivers may have been only one of several causes of the drop in pCO_2 at that time, which contributed to the start of major ice accumulation in the Northern Hemisphere, and that accumulation may in turn have been one of several factors influencing the East Asian monsoon.

19.6 Conclusion

The collision of Greater India and Asia some 45 million years ago caused uplift of the Tibetan Plateau and led to major changes in atmospheric circulation and climate across Asia. Evidence of early cooling associated with this uplift dates back to 38 Ma in China, with marked cooling at the transition from Eocene to Oligocene 34 Ma ago, when permanent ice became established on Antarctica. Shrinking of the Tethys Sea during the Oligocene and Miocene accentuated the growing aridity in Asia, with desert dust deposition by at least 24–22 Ma. The monsoon system developed in stages, with strong rainfall seasonality by 10.7 Ma. The unique Chinese loess record spans the entire Quaternary and consists of alternating couplets of unweathered loess and fossil soils. The soils developed during warm, wet interglacial and interstadial phases when the Asian summer monsoon was more active, and the loess accumulated during cold, dry glacial and stadial phases when the summer monsoon was weak and the winter monsoon was strong. The fragmentary record from dunes, lakes and rivers is in general consistent with the loess record, although in the Thar Desert of India (and the Wahiba Sands of Arabia), dune accretion was greatest at 15–14 ka, when the summer monsoon was once more becoming strong. In this case, wind velocity outweighed aridity as a cause of dune mobility. Elsewhere, the LGM was cold, dry and windy, with dune movement and dust mobilisation. Speleothem samples obtained from caves in China provide a high-resolution record of monsoon history for the past 200 ka. It appears that when the Greenland ice cores indicate a warmer regional climate, the

Asian monsoon is stronger, and conversely. One unprecedented event was the 74 ka eruption of Toba volcano in Sumatra, which also left India covered in a blanket of volcanic ash 10–15 cm deep. The climatic impact of this eruption, which was the largest eruption of at least the last 1 million years, is still in dispute. Pollen and carbon isotopic evidence suggests a change in plant cover over north-central India, with forest prevalent before and grassland and open woodland prevalent after the eruption. It is still not clear what effect this event had on humans and other organisms away from the immediate site of the eruption.

20

North American deserts

Climb the cliff at the end of Labyrinth Cañon, and look over the plain below, and you see vast numbers of buttes scattered about over scores of miles, and every butte so regular and beautiful that you can hardly cast aside the belief that they are works of Titanic art.

John Wesley Powell (1834–1902)

Explorations of the Colorado River of the West (1875)

20.1 Introduction

Although the individual deserts in North America and northern Mexico are quite small when compared to the great deserts of the Sahara, Arabia, Australia and Asia, no other continent has had such an enormous impact on global climatic changes throughout the Quaternary as North America. Great ice caps waxed and waned in the north of the continent and had a profound influence on global sea level and global oceanic and atmospheric circulation (see [Chapter 3](#)). The most recent of these was the Laurentide ice cap – so-named because it was located over the Laurentian shield. This ice cap was up to 3 km thick and reached its maximum extent as recently as 20,000 years ago. As a result of the vast amounts of water locked up in the ice caps of North America, Greenland and north-west Europe during the Last Glacial Maximum, global sea levels fell by 120 m, exposing hitherto submerged continental shelves (Chappell, 1974; Chappell et al., 1996; Lambeck and Chappell, 2001). Air temperatures to the south and west of the Laurentide ice cap were so low throughout the year that the ground was permanently frozen except along the southern margins of the permafrost region, which thawed out briefly at the surface during the short summers. It was not until the ice caps began to melt and a corridor opened up between the shrinking Cordilleran ice cap to the west and the Laurentide ice cap to the east that prehistoric hunters were able to enter North America from eastern Siberia across the Bering land bridge and down the Mackenzie Valley some 14,000 years ago, although some



Figure 20.1. Horizontal bedding of rocks exposed by erosion, Grand Canyon, Arizona.

earlier arrivals by other coastal routes cannot be ruled out. Whether the arrival of these hunters was the primary cause of the late Pleistocene and early Holocene extinctions of the North American megafauna is still a matter of vigorous debate (see [Chapter 17](#)).

The aim of this chapter is to describe the environmental fluctuations experienced by the drier parts of North America, bearing in mind that many areas that are now relatively humid were previously arid, just as many of the arid areas were once far wetter than they are today.

20.2 Desert landscapes of North America

The desert landscapes of North America are among the best-known landscapes on earth (Thornbury, 1965; Graf, 1987b) ([Figures 20.1 to 20.3](#)). The detailed three-dimensional sketches of John Wesley Powell, Grove Karl Gilbert and many other intrepid geological explorers of the mid-nineteenth century brought to public attention the monumental grandeur and mystery of such features as the Grand Canyon (see [Chapter 5](#)). Attempts to explain the origin of these unique landforms resulted in the formulation of some of the most fundamental concepts and principles of geomorphology, including notions of river base level, stream antecedence, superposition, competence and capacity (see [Chapter 10](#); Baulig, 1950; Chorley et al., 1964; Sparks,



Figure 20.2. Amphitheatre headwall created by differential erosion, Grand Canyon, Arizona.

1972; Tchakerian, 1995). The role of the sporadic desert sheet floods and of running water more generally was quickly recognised in even the most arid of landscapes, leading to the recognition of ancient river systems on the Colorado Plateau and elsewhere (Luchitta et al., 2011). Then came the recognition of a suite of desert landforms, including pediments, or low-angle rock-cut slopes, at the foot of mountains in Arizona, alluvial fans in the Mojave Desert and salt lakes, or playas, in Utah.

Efforts to account for the differences between landforms deemed typical of humid regions and those thought to be characteristic of deserts led to prolonged debates over whether slopes in deserts retreated parallel to themselves to form low-angle slopes of coalescent pediments termed pediplains or whether they were worn down



Figure 20.3. Desert landscape, Arizona.

to produce peneplains of the type espoused by William Morris Davis (1909; 1912) and later popularised by the great New Zealand geomorphologist Sir Charles Cotton (1947), who considered desert landscapes to be ‘climatic accidents’ from the so-called ‘normal cycle of erosion’. Such debates ultimately proved sterile, given that the opposing conclusions attained by the different advocates are in fact scientifically untestable and arise simply from the initial unproven assumptions implicit in each proposed model of landscape evolution. Be that as it may, the public and scientific focus on desert landscapes prompted careful attention to the processes of weathering and erosion in arid regions and the recognition that present-day desert landscapes are polygenic, having been fashioned over many millions of years under a variety of different climates (Mabbutt, 1977; Frostick and Reid, 1987a; Cooke et al., 1993; Thomas, 1997; Parsons and Abrahams, 2009). With these general ideas in mind, let us now consider where the North American deserts fit into the major landscapes of that continent.

Put very simply, North America consists of three main geomorphic elements: the western mountains, the central lowlands and the eastern hills (Figure 20.4). Each may be subdivided into distinct smaller units (Thornbury, 1965; Graf, 1987b). From west to east, the western mountains include the Coastal Ranges and Sierra Nevada, the High Plateaux, and the Northern and Southern Rockies. The High Plateaux include the Lava Plains in the north, the Great Basin in the centre, and the Colorado Plateau

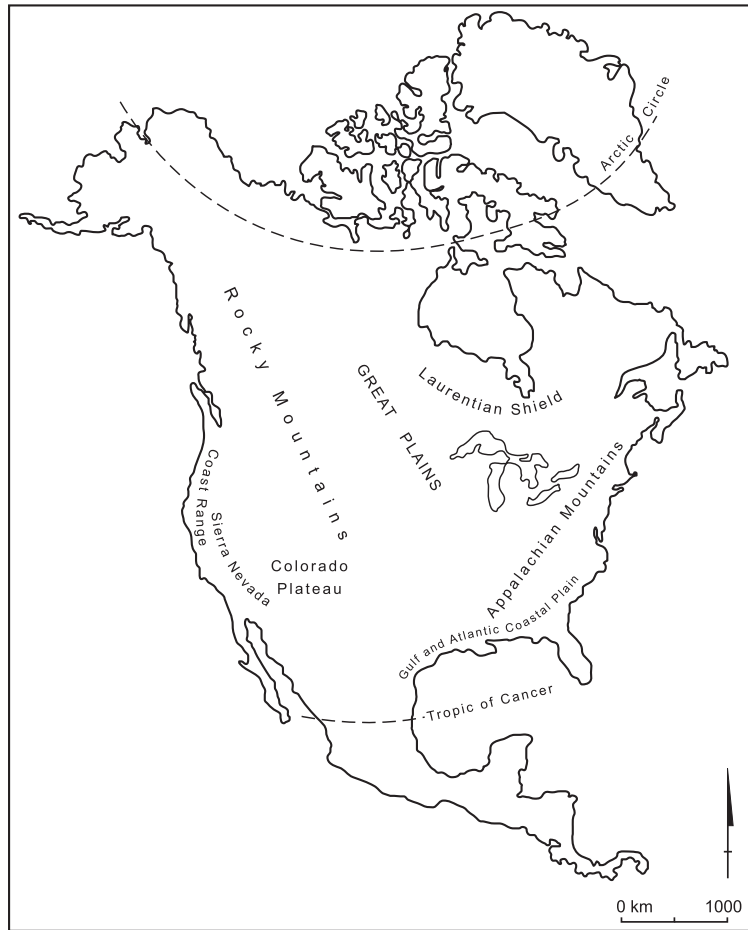


Figure 20.4. Major physiographic regions of North America.

and Basin and Range Province in the south. The central lowlands are made up of the Laurentian or Canadian Shield to the north, the Great Plains and the Central Lowland to the south, and the Gulf and Atlantic Coastal Plain in the far south. The eastern hills include the Appalachians in the west, the Piedmont region in the south and the isolated Ozarks to the east. The deserts and drier areas coincide very broadly with the Colorado Plateau, the Basin and Range Province, and the Great Plains.

The Rocky Mountains are the single most important physiographic feature in North America in terms of their effect on climate. They extend for 5,000 km from the tropics to the subarctic and are aligned north-south, perpendicular to the dominant westerly winds. Although not as high as the mountains of Asia or the Andes, the Rockies are high enough to have a permanent cover of snow in the north, where they reach a maximum elevation of 6,196 m (20,320 feet) on Mount McKinley in Alaska. Two

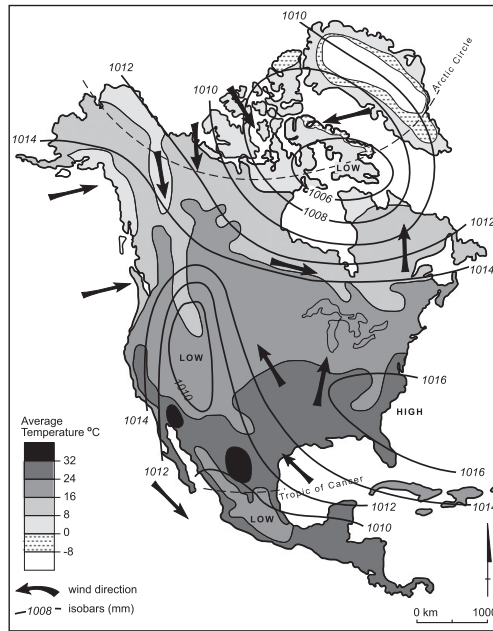
smaller ranges, the Cascades and the Sierra Nevada, run roughly parallel to the main Rockies.

The two largest desert areas are the Chihuahuan Desert (450,000 km²), which extends from north-central Mexico into southern New Mexico, and the Sonoran Desert (300,000 km²), which extends from north-west Mexico into southern Arizona. The Sonoran Desert is bounded to the east by the Sierra Madre Occidental, and receives most of its rain from the Pacific during the summer. The Chihuahuan Desert lies between the Sierra Madre Occidental and the Sierra Madre Oriental, with most of its rainfall in summer, supplemented by sporadic tropical cyclones from the Caribbean. The smaller Mojave Desert (140,000 km²) lies north-east of Los Angeles and south-west of Las Vegas, within the rain shadow of the Transverse Ranges. Precipitation occurs mainly in winter, with snow on the higher ranges. The Colorado Plateau (375,000 km²) and the Great Basin (410,000 km²) are also arid, although their higher elevation means that they are slightly cooler in summer than the less-elevated southern deserts. The Baja California peninsula is also very dry and extends slightly south of the Tropic of Cancer. We now consider the causes of aridity in North America and northern Mexico.

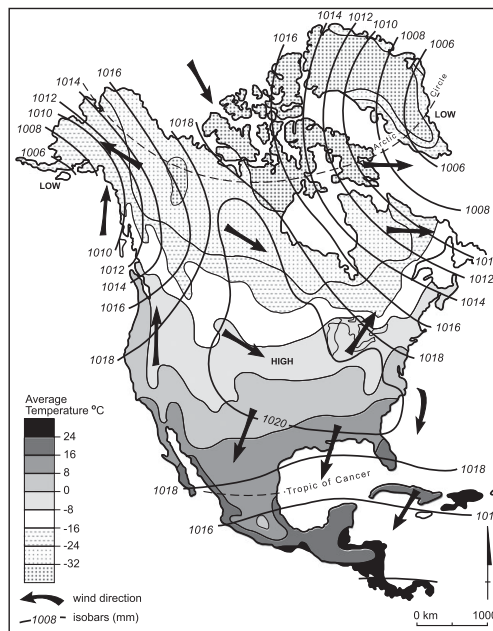
20.3 Present-day climate and causes of aridity

Much of North America lies within the zone of the westerlies, but the Rockies are aligned perpendicular to these moist air masses and provide a formidable barrier to their eastward passage. As a consequence, the regions east of the Rockies are in a vast rain shadow, extending in the north from North Dakota through South Dakota, Nebraska, Kansas and Oklahoma to Texas in the south. The Great Basin of Nevada and much of Utah, the Colorado Plateau and Arizona also lie in the rain shadow of the Rockies. The Great Salt Lake Desert of Utah is an eloquent witness to this aridity. The climate of much of inland North America is best described as continental, with hot, dry summers and very cold winters. The absence of any east-west aligned mountain barriers also means that outbreaks of polar air can penetrate far to the south in winter, bringing snow and extreme cold as far south as Florida, Louisiana and Texas (Kendrew, 1961, p. 412, fig. 127).

The west coast lies in the direct path of the westerlies, which bring rain throughout the year to Oregon and Washington and winter rain to California (Figure 20.5). An additional factor accentuating the summer aridity of California and Baja California is the presence offshore of the cold California Current, which brings coastal fog to Los Angeles and San Francisco. The Sonoran and Chihuahuan Deserts and the southern tip of Baja California lie beneath the descending arm of the tropical Hadley Cell, which serves to intensify their aridity (see Chapter 1). Because the prevailing winds are blowing offshore along the east coast, the zone of high coastal rain is relatively narrow. An exception is the well-watered region around the Gulf of Mexico, which



(a)



(b)

Figure 20.5. Surface wind and pressure patterns (a) during the northern summer (July) and (b) during the northern winter (January). (After *The Times Atlas of Americas*, 2010.)

receives warm, moist air from both the Atlantic and Pacific, as well as torrential rains from tropical cyclones, which can extend well inland to Oklahoma and beyond.

20.4 Mesozoic and Cenozoic tectonism, cooling and desiccation in North America

Several seemingly unrelated factors were responsible for the origin of the arid and semi-arid regions of North America and northern Mexico. First and most important were the tectonic movements that gave rise to the Cordillera or Rocky Mountains in the west, the Basin and Range Province, the Colorado Plateau and the Great Plains. Next in importance was the progressive cooling of the Arctic, which culminated in the expansion of sea-ice and the accumulation of land-ice across the Laurentian shield region of North America. As temperatures fell and the climate became drier during the late Cenozoic, forests gave way to grasslands and the fauna changed accordingly. Superimposed on these long-term cooling and drying trends were the relatively rapid climatic fluctuations associated with the waxing and waning of the great ice sheets of North America and Greenland. The arrival of periodic pulses of meltwater into the North Atlantic via the St. Lawrence River and down the Mississippi River into the Gulf of Mexico caused changes in sea surface temperatures and local climate. More fundamental were alterations to the deep-water circulation in the North Atlantic and concomitant changes in what has been termed the global ‘thermohaline’ oceanic circulation system (Broecker, 1992; Williams et al., 1998), leading on occasion to severe droughts in North Africa and the Near East. The adjective ‘thermohaline’ simply means that the ocean currents are driven by density differences linked to temperature and salt content, with warm freshwater being less dense than cold, salty water. The denser water sinks and becomes part of the deep-water circulation, while the less dense water remains at the surface.

The tectonic history of North America is complex and has been reviewed in detail elsewhere. For comprehensive summaries of the earlier work, see Bally and Palmer (1989) and Bally et al. (1989). Our aim here is simply to summarise the major tectonic events relevant to the inception of the present-day arid lands of North America. The origin of the Rockies and of the Colorado Plateau and the Basin and Range Region is tied up with the history of the Farallon oceanic plate, which began to move beneath the west coast of North America during the Jurassic. The western margin of this plate coincides with the East Pacific Rise (see [Chapter 3](#)).

In a widely cited paper, Dickinson and Snyder (1979) argued that once the San Andreas transform fault had developed, thereby separating the Farallon plate from the North American plate, the lack of subduction at the transform plate boundary led to the growth of a slab-free zone beneath the continental block near the San Andreas transform. They went on to argue that diapiric upwelling of magma from the asthenosphere led to uplift of the adjacent Sierra Nevada and Colorado Plateau and the

opening of the Rio Grande Rift. Later workers have disputed this interpretation, with Schellart et al. (2010) offering an alternative interpretation involving the Cenozoic slowdown of the Farallon plate and the ensuing decrease in the rate of subduction. They concluded that the change from the Sevier-Laramide orogenic regime to the Basin and Range extensional tectonic regime was related to the width of the subducting Farallon slab in western North America. Uplift and deformation in the Rocky Mountains began in the Jurassic some 160 Ma ago, culminating in uplift of the Sevier Mountains and Canadian Rockies around 125 Ma ago. Schellart et al. (2010) postulated that these orogenic events were associated with the subduction of a wide slab and crustal shortening, whereas extension began in the Eocene around 45 Ma ago when there was a phase of intermediate to narrow slab extension and slab-driven trench retreat.

Regardless of the precise causes of uplift and extension, our primary concern here is with the timing and amount of uplift or subsidence, which is often surprisingly hard to establish from existing data. Determining how and when the Grand Canyon formed illustrates this point well. There are several schools of thought in regard to what Rebecca Flowers (2010) has called ‘the enigmatic rise of the Colorado Plateau’ and the equally enigmatic history of the Colorado River and its incision into the Colorado Plateau (Figures 20.1 and 20.2). One view is that uplift of the plateau and corresponding river incision is geologically recent, possibly no more than about 5 million years. The opposing view is that uplift began at least 25 million years ago. An arsenal of highly ingenious techniques has been used in an effort to solve the enigma.

Sahagian et al. (2002) observed that the size of the uppermost vesicles within basalt flows reflects surface atmospheric pressure at the time that they formed, and they postulated that the inferred atmospheric pressure can then be used to estimate the surface elevation at the time when the lava flows were laid down. Most of the lava flows in the Colorado Plateau were emplaced over the past 25 Ma at what are now elevations of 800–2,000 m. The results of the vesicle analysis showed that uplift was already underway by 25 Ma, with 800 m of uplift before 5 Ma at an average rate of 40 m/Ma and 1,100 m of uplift after 5 Ma at a mean rate of 220 m/Ma. They concluded that initial Miocene uplift was relatively slow, with an order of magnitude increase in the rate of uplift during the Pliocene and Quaternary.

Polyak et al. (2008) used a special type of speleothem called ‘cave mammillaries’ to determine past elevations of the water-table at nine sites in the Grand Canyon. They obtained uranium-lead ages for the speleothems and assumed that incision into the Grand Canyon was accompanied by a fall in the local groundwater-table, which seems reasonable. They further assumed that the rate of incision was comparable to the rate of water-table fall. On this basis, they concluded that incision in the western Grand Canyon amounted to 55–123 m/Ma over the past 17 Ma, while incision was more rapid in the eastern Grand Canyon, amounting to 166–411 m/Ma. They went on to propose that the Grand Canyon was formed by headward erosion from west to east, with an acceleration in incision in the eastern sector during the past 3.7 Ma or so.

Liu and Gurnis (2010) proposed a model of uplift based on Farallon slab evolution and mantle convection to account for uplift of the south-west plateau. They concluded that up to 1,200 m of uplift took place between the Late Cretaceous and the Eocene, with an initial phase of subsidence followed by uplift in two stages, one in the Late Cretaceous, the other in the Eocene. An apparently opposing model based on late Cenozoic edge-driven convection along the plateau margins indicated several hundred metres of late Cenozoic uplift (van Wijk et al., 2010). In fact, as Flowers (2010) noted in her commentary on these papers, the two models are not mutually exclusive, and both early and later uplift are possible. Flowers and Farley (2012) later obtained apatite $^4\text{He}/^3\text{He}$ and (U-Th)/He thermochronometric ages for both the eastern and western Grand Canyon and concluded that the western Grand Canyon had been incised to within a few hundred metres of its modern depths by around 70 Ma ago, refuting the view that the entire canyon was cut in the last 5 to 6 Ma. It is even possible that the ancestral Colorado River flowed north-east into the Labrador Sea during the late Oligocene-early Miocene before its capture by a river flowing to the Gulf of California (Sears, 2013).

Hyndman and Currie (2011) observed that the crust beneath the North American Cordillera is only 30–35 km thick, in contrast to the 40–45 km of the low-elevation craton and other tectonically stable areas to the east of the Cordillera. They concluded that the Cordillera is located in a former back-arc zone where temperatures are uniformly hot, with thermal expansion responsible for about 1,600 m of the present Cordillera elevation compared to the colder stable areas. However, uplift was not synchronous along the entire length of the North American Cordillera. Mix et al. (2011) examined around 3,000 stable isotope measurements that had been used as an index of surface elevation and concluded that Eocene uplift propagated from north to south, leading to the development of an Eocene-Oligocene highland 3–4 km in elevation, by which time modern patterns of precipitation had been established. Although there was continued uplift, faulting and local subsidence after that time, continuing in to the Quaternary, the major topographic elements governing the location of the present-day arid and semi-arid areas in North America and northern Mexico were well-established by the late Oligocene. Before proceeding to consider how these tectonic events influenced the biota in the deserts and semi-deserts of the region, we need first to consider Cenozoic climatic changes in the Arctic and their impact on North America.

20.5 Cenozoic cooling of the Arctic

The early Cenozoic environment of the Arctic region was very different from that of today (Grantz et al., 1990), culminating in the Palaeocene-Eocene thermal maximum of around 55 Ma, when a brief period of rapid global warming took place at a time when the world was already warm, most probably as a result of an increase in atmospheric greenhouse gas concentrations (Pagani et al., 2006; Sluijs et al., 2006).

At that time, sea surface temperatures near the North Pole increased from about 18°C to more than 23°C (Sluijs et al., 2006). The transition from the Palaeocene and early Eocene warm ‘greenhouse’ world to a colder world with sea-ice and icebergs in the Arctic Ocean was underway from middle Eocene times onwards (Moran et al., 2006). Marine sediment cores from the Lomonosov ridge in the Arctic Ocean show the first evidence of ice-rafted debris in the middle Eocene 45 Ma ago, with fresh surface waters present by around 49 Ma (Moran et al., 2006). The earliest cooling in the Arctic is coeval with that in the Antarctic, suggesting that global rather than regional factors were responsible for the initial cooling. Eldrett et al. (2007) found abundant evidence of deposition of macroscopic ice-rafted debris in the Norwegian-Greenland Sea sediments between 38 and 30 Ma, and concluded that isolated glaciers were probably present on Greenland in the late Eocene to early Oligocene. The presence of the freshwater fern *Azolla* in Eocene Arctic Ocean waters is also indicative of pulses of fresh surface water discharge from the Arctic Ocean to the south, some as far as the North Sea (Brinkhuis et al., 2006).

Stickley et al. (2009) found abundant sea-ice dependent fossil diatoms (*Synedropsis* spp.) in middle Eocene marine sediments from the Arctic Ocean. They concluded that the establishment of sea-ice in this region took place in two stages: an initial phase of sporadic sea-ice formation in marginal shelf areas starting around 47.5 Ma ago, followed 0.5 Ma later by seasonal sea-ice formation offshore in the central Arctic. They stressed the need to distinguish between sea-ice formation and evidence for ice-rafted debris from the land, noting that sea-ice affects ocean-atmosphere exchanges, while land-ice affects sea level and ocean acidity.

Migrations of Arctic marine organisms (dinoflagellates) and changes in the $\delta^{18}\text{O}$ composition of benthic forams indicate strong mid-Oligocene cooling of surface water in the Arctic at around 27.1 Ma, synchronous with Antarctic ice sheet growth and sea level lowering (Van Simaeys et al., 2005). The opening of Fram Strait in the early Miocene allowed the free flow of deep water from the Arctic to the North Atlantic and the inception of the global thermohaline circulation system (Jakobsson et al., 2007). This was a precursor to the progressive late Pliocene cooling of the North Atlantic (Loubere, 1988), which culminated in the high-latitude accumulation of snow and ice in North America and the onset of major North American glaciation at around 2.5 Ma. The Pliocene cooling was relatively rapid; Early Pliocene fossil plants, mammal bones and beetles on Ellesmere Island in northern Canada indicate that summer temperatures were 10°C higher and winter temperatures were 15°C higher than they are today (Tedford and Harington, 2003).

20.6 Cenozoic desiccation of North America and its impact on the biota

Flannery (2001) has provided a useful overview of changes in the Cenozoic flora and fauna of North America caused by changes in regional climate and topography, as

well as in patterns of migration and interspecies competition. To provide a detailed account of these topics is beyond the scope of the present work, so only a few of the more salient topics will be considered here.

The Palaeocene-Eocene thermal maximum (around 55.8 Ma), recorded in the Arctic and many other global marine records, is also evident on land, but sites with fossils of this age are quite rare. Two sites from the northern Rockies in Wyoming have yielded plant macrofossils (see [Chapter 16](#)) from alluvial deposits spanning the Palaeocene-Eocene boundary (Wing et al., 2005). These unique floras are a mixture of native North American species from the south and east and immigrant species from Europe. The $\delta^{13}\text{C}$ composition of the organic matter in soils developed on fine-grained alluvium indicates that temperatures rose by about 5°C. Leaf area analysis suggests that mean annual rainfall dropped by approximately 40 per cent at the start of this interval and then increased late in the interval, when the warm, wet climate allowed the formation of thick paleosols (Wing et al., 2005). The response of the flora to these climatic fluctuations was species specific and relatively rapid, taking place within about 10,000 years, which is comparable to the postglacial changes in the North American flora described in [Section 20.7.5](#).

As noted in [Section 20.5](#), there was a sudden drop in temperature at the Eocene-Oligocene transition around 33.5 Ma ago, both in the Arctic and in the oceans throughout the world. Less well-known is the degree and impact of this cooling on land. Zanazzi et al. (2007) investigated the $\delta^{18}\text{O}$ composition of fossil tooth enamel and fossil bone spanning this transition at six localities scattered across Wyoming, South Dakota and Nebraska. They found a drop in mean annual temperature of $8.2 \pm 3.1^\circ\text{C}$ over a period of about 400,000 years and a possible increase in temperature seasonality, but no perceptible evidence of any change in aridity at that time, from which they concluded that the drop in temperature was caused by a significant decrease in the atmospheric concentration of carbon dioxide.

The expansion of grasslands across the world took place in the late Cenozoic and appears to be related to changes in temperature, precipitation and, perhaps, atmospheric carbon dioxide concentration. In the Great Plains, a series of reasonably well-dated fossil soils allows us to follow the change from forests to grasslands over the past 23 Ma. Fox and Koch (2003) analysed the stable carbon isotopes in the soil carbon and soil carbonate in a series of paleosols in the Great Plains extending from Nebraska and Kansas through New Mexico and Oklahoma to Texas. Plants that follow the C_3 photosynthetic pathway ([Chapter 7](#)) (henceforth, C_3 plants) have present-day mean $\delta^{13}\text{C}$ values of -27‰ with a range of -35‰ to -22‰ . C_4 grasses have mean $\delta^{13}\text{C}$ values of -13‰ , with a range of -14‰ to -10‰ , and comprise 50–100 per cent of the modern biomass in this region. Fox and Koch (2003) found that the C_4 biomass amounted to 12–34 per cent during the Miocene, increased between 6.4 and 4.0 Ma and attained present levels by 2.5 Ma. They also noted that the Miocene ecosystems of the Great Plains have no modern analogue and concluded that the

Miocene and Pliocene changes in both flora and fauna on the Great Plains reflected local and regional factors at least as much as they reflected more global climatic fluctuations.

20.7 Quaternary climatic fluctuations

20.7.1 Glaciations

Progressive cooling in the Arctic during the late Pliocene resulted in rapid accumulation of snow and ice and the inception of a large ice sheet over the North American Laurentian shield by around 2.5–2.6 Ma ago. Much of the terrestrial evidence of the older glaciations has long since been eroded and destroyed, so inferences about the magnitude and duration of past glaciations is indirect and comes from the marine sediment record, especially the oxygen isotopic composition of the calcareous shells of foraminifera. The early Pleistocene ice caps were smaller than their late Pleistocene counterparts, and the duration of the glacial-interglacial cycles was also not as long. Deep-sea sediment cores show that the 23 and 19 ka precessional cycles (see [Chapter 3](#)) were dominant until around 2.6 Ma. Between about 2.5 and about 0.7 Ma, the 41 ka obliquity cycle was dominant, with the 100 ka orbital eccentricity cycle dominant thereafter (Raymo et al., [1997](#); Williams et al., [1998](#); Lisiecki and Raymo, [2005](#); Clark et al., [2006](#); Lisiecki and Raymo, [2007](#); Lüthi et al., [2008](#)).

Detailed field mapping of glacial deposits and intercalated warm-climate soils and their associated pollen spectra during the late nineteenth century led to the recognition of four major Pleistocene glacial stages and three major interglacial stages in North America. From oldest to youngest, these were the Nebraskan glacial, the Aftonian interglacial, the Kansan glacial, the Yarmouth interglacial, the Illinoian glacial, the Sangamon interglacial and the Wisconsin glacial (Flint, [1971](#), p. 543, table 21-B). At about the same time, Penck and Brückner ([1909](#)) identified four recent glacial stages in the piedmont zone of the European Alps, which from oldest to youngest they termed the Günz, Mindel, Riss and Würm glacial stages. The corresponding interglacial stages can be labelled quite simply as the G/M, M/R and R/W interglacial stages. Later work showed that this fourfold division was oversimplified, because many of the deposits associated with each glacial were in fact polygenic and represented more than one glacial stage (see Williams et al., [1998](#), for a review of this topic). Two much older sets of glacial deposits were also recognised, and they were termed the Biber and the Donau glacial stages. Older and highly weathered glacial deposits were likewise identified in North America well beyond the limits of the last, or Wisconsin, glaciation. The Wisconsin glacial deposits were later subdivided on the basis of stratigraphy and radiocarbon ages into Early (>55,000 years BP), Middle (55,000–25,000 years BP) and Late Wisconsin (25,000–10,000 years BP) (Flint, [1971](#), p. 560).

The *relative* sequence of glacial and interglacial events in North America is now quite solidly established, and where glacial deposits are lacking, use is made of the

distinctive loess deposits (see [Chapter 9](#)) downwind of the ice caps and blown from the glacial outwash plains. As in China, the loess deposits often consist of a lower, relatively unweathered unit and an upper, often highly weathered unit in which a soil had developed. However, despite the establishment of an impressive relative sequence of glacial and interglacial stages, there was no way of knowing how much of the North American continental record was missing until the deep-sea sediment records across the oceans of the world revealed that there had been more than thirty glacial stages in the last 2 million years. How the Marine Isotope Stages (MIS) were determined is explained in [Chapter 6](#). Each unevenly numbered one represents an interglacial or an interstadial stage, with the present-day interglacial stage designated as MIS 1. All counting is from the present back in time. Each evenly numbered marine isotope stage represents a glacial or a stadial stage, with the Last Glacial Maximum (LGM) designated as MIS 2. Further subdivision of each stage is shown by letters of the alphabet, with, for example, the peak of the last interglacial designated as MIS 5e. MIS 3 is an interstadial, not a full interglacial, which is confusing for the unwary.

The vast size and high elevation of the Laurentide ice sheet (coeval with the Late Wisconsin glacial stage) and, no doubt, its earlier equivalents, may have caused the high-level jet streams to split and deviate from their interglacial positions over North America, thus altering the lower-level patterns of circulation over the continent (Kutzbach and Wright, 1985). However, more recent work suggests that these changes were unlikely to have caused any major southward displacement of the westerlies during the LGM (Lyle et al., 2012).

Relative to the ice caps, mountain glaciers showed a far more variable response to late Pleistocene climatic change. This is to be expected, given that glaciers will respond to local differences in valley size, shape and aspect, not to mention local influences on precipitation. In addition, until the advent of cosmogenic nuclide dating (see [Chapter 6](#)) allowed glacial moraines to be directly dated for the first time, all previous late Quaternary chronologies relied on radiocarbon dating of organic materials sandwiched between glacial deposits and were therefore at best indirect measures of the timing of glacial advances and retreats. Porter et al. (1983) have summarised much of the earlier work on Late Wisconsin mountain glaciation in the western United States, which was based primarily on ^{14}C dating. Since that time, the widespread deployment of cosmogenic exposure dating has confirmed earlier work while also adding significant new detail. Indeed, the late Pleistocene and Holocene glacial history of the North American deserts is without a doubt the best-dated of any desert region on earth (Porter, 1983; Wright, 1984; Ruddiman and Wright, 1987; Wright et al., 1993; Easterbrook, 2003; Clark et al., 2009; Young et al., 2011).

Many of the mountain glaciers in the drier regions of North America were at or near their maximum limits during the Last Glacial Maximum, but there was still significant local variation. The ^{10}Be surface exposure ages from moraines, bedrock and river terrace gravels relating to the late Pleistocene Pinedale glaciation from three adjacent

valleys in the upper Arkansas River Basin illustrate how local topography can override regional climate (Young et al., 2011). Maximum glacier advances in these three valleys were far from synchronous, ranging from 22 to 16 ka, although glacier retreat began at 16–15 ka in all three valleys. The ages obtained for other Pinedale glacial moraines at widely different sites in the western United States show that although climatic change exerts a major influence on glacier advances and retreats, local factors such as aspect and topography modulate this influence (Young et al., 2011).

One long-standing issue in glacial geology is whether the late Pleistocene glacial advances and retreats were synchronous or out of phase in both hemispheres. In an effort to resolve this question, Clark et al. (2009) analysed 5,704 ^{14}C , ^{10}Be and ^3He ages from across the globe. They found that the Northern Hemisphere ice sheets (Greenland, North America, Europe and parts of Asia) began to retreat at roughly the same time (20–19 ka) regardless of size, as did most Northern Hemisphere mountain glaciers. However, the mountain glaciers of Tibet and those of the Southern Hemisphere started to retreat somewhat later (18–16 ka), and the West Antarctic Ice Sheet retreated later still (around 14.5 ka). Mountain glaciers in many areas were already at or near their maximum extent by around 30 ka, which was when global ice sheets began to reach their maxima. The period of minimum global sea level (26.5–19 ka) was when the global ice sheets were in near equilibrium with climate. According to Clark et al. (2009), melting of the northern ice sheets was likely related to three main forcing factors: an increase in high northern latitude insolation, an increase in atmospheric CO_2 concentration and a rise in tropical Pacific sea surface temperatures.

Another as yet unsolved problem in regard to Quaternary climates is why the lead up to glaciation involved such a long interval of cooling, whereas the melting of the ice caps was the result of a relatively short phase of warming. Denton et al. (2010) considered this question, noting that the postglacial melting of the Northern Hemisphere ice caps to present volumes ‘represents one of the largest and most rapid natural climatic changes in Earth’s recent history’ (op. cit., p. 1652). During this rapid warming and melting, sea level rose by 120 m and atmospheric CO_2 increased by 100 ppmv. The influx of meltwater into the North Atlantic led to cold conditions in the Northern Hemisphere and a change in oceanic and associated atmospheric circulation patterns. They speculated that during each northern cold stadial event, the Southern Hemisphere westerlies shifted to the south, resulting in warming and deglaciation in the Antarctic and in the Southern Ocean.

20.7.2 Loess, dunes and fossil soils

The loess deposits of North America were not always recognised as eolian dust, and Darwin’s contemporary, the great geologist Charles Lyell, was convinced that the loess along the Mississippi Valley was alluvial in origin (see Chapter 9). However, by the late nineteenth century, the reality of widespread Quaternary glaciations had

been accepted, and detailed field mapping revealed that the loess was associated with deflation from the outwash plains along the southern margins of the Laurentide and Cordilleran ice caps. The loess is not necessarily synchronous with full glacial conditions only, because it continued to form during early deglacial times, which were still cold and windy and saw limited plant cover. Luminescence dating of loess has enabled the rates of loess accumulation to be determined with increasing accuracy and precision. For example, the rate of accumulation of loess in what is now semi-arid Nebraska was very high between 18 and 14 ka (Roberts et al., 2003). The high atmospheric dust loading over that area may have caused the climate there to remain colder than present for several thousand years, despite higher-than-present summer insolation values (Roberts et al., 2003). Away from the ice margins, in the deserts of the south-west, wind-blown dust derived from bare alluvial surfaces and small patches of sand dunes contributed fine material in the form of clay and calcium carbonate to the soils, which began to form as the climate became less arid and plants colonised the previously bare surfaces of alluvial fans and river flood plains. Much as in central Asia, the loess deposits of the Great Plains consist of alternating units of loess and fossil soils (see Chapter 9). As a general rule, the loess and desert dust accumulated during colder, drier, windier episodes when the glacial outwash and desert source areas were more extensive and frost action was pronounced across the landscape (Péwé, 1981; Pye, 1987; Maher et al., 2010). The soils developed on the loess reflect a warmer, wetter climate and an established plant cover. This applies equally to the alluvial fans and pediments of the Basin and Range Province, where the late Pleistocene soils are relatively thick and have well-developed soil horizons (see Chapter 15), in contrast to the much thinner and less-developed Holocene soils (McFadden et al., 1986; Wells et al., 1987; Dohrenwend et al., 1991). Soils overlying volcanic rock often contain eolian quartz indicating that the parent material was blown in during dust-storms.

In the Chihuahuan, Sonoran and Mojave deserts, the area covered by desert dunes is quite limited. The gypsum dunes in White Sands National Monument in the Chihuahuan Desert were blown from gypseous playa deposits, and often form a series of crescentic dunes reminiscent of the gypseous lunettes of Algeria and southern Australia. In the Mojave Desert, many of the sand ramps deposited against the foot of the mountains are polygenic and contain fossil soils as well as colluvial and alluvial deposits (Tchakerian, 2009). The *Gran Desierto del Altar* sand sea (5,700 km²) in the north-west Sonoran Desert is at the distal end of the Colorado River and has been subject to repeated changes in local climate, tectonic events and changes in the location of the Colorado River delta, with the result that the dunes have varied orientations and often cross older, underlying dunes (Warren, 2013, fig. 8.12).

The Nebraska Sand Hills (51,000 km²) are the most extensive of the former sand seas on the High Plains, and are continued to the west and south by smaller dune fields in Wyoming, Colorado, Kansas, New Mexico, Oklahoma and Texas. The majority of these dunes are now vegetated and stable, although in extreme drought years they

can become reactivated. The dominant wind direction when these dunes were active was from the north-west during times when the large ice caps over North America may have partly diverted the westerlies (Kutzbach and Wright, 1985). Factors contributing to sand movement during glacial times include extreme cold, strong winds, depleted or absent vegetation and a supply of available sand from older dunes. The low concentration of atmospheric carbon dioxide during glacial maxima, amounting to 180–200 ppmv, would have also contributed to the sparse plant cover (Warren, 2013, p. 161). Both the Nebraska Sand Hills and the dunes of the Gran Desierto in north-west Mexico have luminescence ages indicating that they were already active during the Middle Pleistocene some 300,000 years ago (op. cit., p. 163).

20.7.3 River fluctuations

Rivers rising in arid and semi-arid regions have a much more erratic flow regime than rivers in humid regions, with long intervals of low flow punctuated by occasional extreme floods (see Chapter 10). Changes towards a more humid climate resulting in a more complete plant cover can lead to changes in river channel pattern (e.g., from braided to meandering) and in the type of load carried (e.g., from coarse traction load to fine suspended load). As a result, the sediments and fossil soils exposed in alluvial fans and river terraces can provide a rough guide to the type of environment in which they were transported and laid down. For example, during the Late Pleistocene, the rivers of the Atlantic Coastal Plain were well beyond glacial limits, and reveal a change from a previous meandering pattern to a braided pattern that was synchronous with local dune formation (Leigh et al., 2004). The dunes were derived from fluvial channel sands and point bars. In desert environments, it is often necessary also to take into account possible sediment inputs to the rivers from sand dunes and wind-blown dust. The highest sediment yields in the rivers of North America are to be found in semi-arid regions (Langbein and Schumm, 1958), and on a more global scale, a second peak occurs in the seasonally wet or monsoonal tropics (Douglas, 1967). The peak in semi-arid areas occurs because of sparse plant cover and the high erosive impact of the rain. In arid areas, there is not enough rain for sustained run-off and fluvial erosion, and in humid areas, the dense plant cover protects the soil from raindrop impact and soil loss from overland flow (see Chapter 10). In the monsoonal tropics, the highest rates of soil loss occur at the start of the wet season, when heavy convective downpours fall on bare, unprotected soil (see Chapter 10).

The situation becomes more complicated when the rivers are located downstream of large lakes that were formed by glacial damming, because the sudden release of water caused by a breach in the ice dam can have catastrophic effects. In Iceland, such a glacial outburst flood (or *Jökulhlaupt*) can result in peak flows of $>50,000 \text{ m}^3/\text{second}$. The ‘Channeled Scablands’ of Washington were formed by a similar sudden release of water from an ice-dammed lake, Glacial Lake Missoula (Baker, 1978; Baker and

Bunker, 1985; Teller, 1995). Glacial Lake Agassiz was larger in area than all the present Great Lakes combined, and it drained at intervals in the late Pleistocene and early Holocene (Teller, 1995).

Late Pleistocene Lake Bonneville overflowed near Red Rock Pass in Idaho about 14,500 years ago. O'Connor (1993) estimated that peak discharge amounted to $1.0 \times 10^6 \text{ m}^3/\text{second}$ at the Lake Bonneville outlet near the pass. Below the outlet, stream power (see Chapter 10) ranged from $10^1 \text{ watts m}^{-2}$ in ponded reaches to $10^5 \text{ watts m}^{-2}$ in constricted reaches. The magnitude of this flood is shown in the size of the cobbles and boulders that were transported and deposited, which ranged from 10 cm to more than 10 m in diameter.

The arroyos and ephemeral stream channels of the American Southwest have been the focus of nearly a century of detailed investigation (Bryan, 1925a; Bryan, 1925b; Leopold and Miller, 1956; Schumm and Hadley, 1957; Bull, 1964a; Bull, 1964b; Lamarche, 1966; Leopold et al., 1966; Tuan, 1966; Haynes, 1968; Cooke and Reeves, 1976; Graf, 1979; Balling and Wells, 1980; Graf, 1982a; Graf, 1983a; Graf, 1983b; Graf, 1987a; Bull, 1991; Schumm, 1991, pp. 108–119; Bull, 1997; Tucker et al., 2006). Bull (1997) distinguished between arroyos and gullies, with the latter being small and ephemeral and the former being up to 200 km long and up to a century old. In the case of both gullies and arroyos, one of the major issues is why they sometimes cut down and sometimes deposit sediment along their beds. Many factors are involved (Cooke and Reeves, 1976), so that no one explanation will cover all cases. Bull (1997) investigated changes in the balance between stream power and resistance to erosion (see Chapter 10). Because both stream power and sediment transport rate are roughly proportional to stream velocity cubed (Schumm, 1977), any factor that reduces stream velocity will ultimately promote channel aggradation. Bull emphasised the importance of plant cover in minimising erosion and in promoting sedimentation within the ephemeral stream network, a conclusion that was confirmed by the work of Tucker et al. (2006) in the semi-arid rangelands of the Colorado High Plains. If the plant cover began to die, as in time of drought, run-off would become less diffuse and would become concentrated around the headwalls of the arroyo, resulting in incision, further drying out of the soil, continued plant death and continued channel incision. Arroyo bank sections revealed multiple episodes of Holocene incision and sedimentation, with less than a century needed for complete incision along the arroyo but more than five times that long needed for complete aggradation (Bull, 1997). Local hydrologic factors appear to be at least as important as regional climatic changes in controlling Holocene arroyo erosion and sedimentation.

Although they might appear to be non-controversial and intuitively valid, the conclusions of Bull (1997) and Tucker et al. (2006) in regard to the influence of changes in plant cover on sedimentation in arid areas have not gone unchallenged. Antinao and McDonald (2013) investigated four localities at different elevations in the Mojave and northern Sonoran deserts, and found that the onset of sedimentation on alluvial

fans in those areas began well before any changes in plant cover upstream. They also observed that sedimentation could occur during several quite different combinations of vegetation change and concluded that factors such as local storm intensity and changes in the routes taken by water and sediment on hill slopes, rather than plant cover, probably controlled late Pleistocene and Holocene fan aggradation in this region. Another study of alluvial fan activity in the Sonoran Desert, this time in the Muggins Mountains near Yuma, showed that fan aggradation was widespread between 3.2 and 2.3 ka (Bacon et al., 2010). These authors concluded that such aggradation was caused by rapid climate change and more intense El Niño-Southern Oscillation (ENSO) events (see [Chapter 23](#) for a review of ENSO). Curiously, they found no sign of any historic reactivation of the alluvial fan surfaces despite rainfall records showing many above-average precipitation events correlated with ENSO. Presumably the late Holocene was a time of greater climatic extremes.

The examples adduced here show that it is hard to discern a clear climatic signal from changes in river channel incision and aggradation, so each catchment needs to be studied in its own right in order to tease out the influence of purely local factors on stream behaviour.

20.7.4 Lake fluctuations

In the Great Basin, there are remnants of more than 100 former lakes, some of them very large (Tchakerian, 1997). The Great Salt Lake in Utah is the shrunken remnant of Late Pleistocene Lake Bonneville, which covered more than 50,000 km² and was up to 330 m deep (Gilbert, 1890; Flint, 1971). Lake Lahontan in north-west Nevada was somewhat smaller (23,000 km²) and up to 275 m deep (Russell, 1885), and Searle's Lake in California is another of many smaller remnants of once large lakes (Flint, 1971; Smith and Street-Perrott, 1983; Lemons et al., 1996; Madsen et al., 2001). By any standards, these were huge lakes. Gilbert (1890) observed the close spatial association between high lake strandlines and glacial moraines and concluded that the lakes were high during times of maximum glaciation. Other workers outside North America reached similar conclusions during the 1860s with respect to the lakes of the Near East and central Asia, such as the Dead Sea and other great water bodies in central Asia, such as the Aral Sea, the Caspian Sea and Lake Balkhash (Flint, 1971). The notion that these 'pluvial' lakes were synchronous with glacials became very firmly entrenched and persisted for nearly a century (see [Chapter 12](#)). The higher lake levels were attributed to a combination of higher precipitation and reduced evaporation from the lake surface caused by lower temperatures. Smith and Street-Perrott (1983) collated the radiocarbon ages for high lake levels in North America and concluded that most of the lakes in what are now arid and semi-arid regions were relatively high during late glacial times. Attempts to resolve the relative importance of the different factors controlling the lake water balance led to thorough hydrologic studies of individual

lake basins (Leopold, 1951; Antevs, 1954; Reeves, 1965). Studies of late Pleistocene snow-lines in the mountains of New Mexico prompted Leopold (1951) and Antevs (1954) to conclude that pluvial Lake Estancia had been high because of decreased temperatures and evaporation and increased precipitation. Reeves (1965) came to a different conclusion in regard to the pluvial lakes of the Llano Estacado in west Texas. He calculated that the Pleistocene precipitation was similar to that of today in this region and maintained that subtle fluctuations in run-off moderated by temperature fluctuations were responsible for the high pluvial lake levels in west Texas.

Gilbert (1890) identified three major high shorelines around Pleistocene Lake Bonneville, which he called Bonneville (about 1,565 m elevation), Provo (about 1,470 m) and Stansbury (about 1,350 m). Owing to isostatic rebound once the lake had fully drained, these elevations should be regarded as relative elevations only. The Provo shoreline is in fact a composite feature formed during periodic overflows between 14,500 and 12,000 ^{14}C yr BP, after which it fell rapidly to its present levels by around 11,500 ^{14}C yr BP (Godsey et al., 2005). Once an overflow channel has been cut, a lake will not be able to rise above that level again. As a result, older and higher lake levels do not necessarily mean that conditions were wetter during those times unless the basin had remained closed throughout the lake's existence. Reheis (1999) considered this question when she examined a number of lakes in the western Great Basin, which did indeed become progressively smaller from early to late Pleistocene. She calculated that the oldest and highest of these lakes would have required an effective precipitation between 1.2 and 3 times that estimated necessary to sustain the late Pleistocene lakes, including Lake Lahontan. Reheis (1999) concluded that the four deep-lake cycles that she identified in these lakes were coeval with marine oxygen isotope stages (MIS) 16, 12, 6 and 2 and therefore occurred during glacials. Oviatt et al. (1999) reached the same conclusion in their re-examination of a deep lake core.

Given that the lakes in the semi-arid American Southwest were high during MIS 2, was it primarily because of higher precipitation or reduced evaporation? In an effort to answer this widely debated question, Menking et al. (2004) chose to re-examine Lake Estancia in central New Mexico using a series of different water balance and run-off models. They concluded that during the LGM, precipitation may have been twice that of today during brief periods of colder and wetter climate, when annual run-off in the basin may have amounted to 15 per cent of annual precipitation as compared to about 2.4 per cent today.

Lyle et al. (2012) have also re-investigated when and why the Great Basin lakes were full during the late Pleistocene. Using a variety of climate proxies including pollen analysis, they studied the incidence of wet phases on either side of the Sierra Nevada along two north-south transects between latitudes 42°N and 32°N. The western transect followed the California coast and considered vegetation changes that were inferred from pollen analysis and estimates of sea surface temperature that were inferred

from alkenone and microfossil assemblages in marine sediment cores collected along the California coast. The eastern transect ran from New Mexico to Nevada, and it included recently acquired ages for the strandlines of Lake Lahontan in Nevada (38°N–42°N) and Lake Estancia in New Mexico (34.8°N). They found that Lake Estancia attained its highest levels between 24.5 and 15.5 ka, with a sudden drop in level at 18–17 ka. Lake Lahontan was high between 25 and 20 ka, low between 19.2 and 17.2 ka, and reached its highest level between 17.2 and 14.5 ka, which was significantly later than when Lake Estancia attained its highest levels. In addition, the central coast of California was relatively wet between 12.5 and 4.5 ka, which was about 5,000 years after the wet interval in southern California. Lyle et al. (2012) concluded that the main source of precipitation in the Great Basin lakes could not have come from any southward displacement of the westerlies during late glacial times, as earlier workers had assumed, but in fact came from the south, specifically from the Gulf of Mexico and the eastern Pacific during the summer monsoon season. In short, the onset of the wettest climate was diachronous in the Great Basin, and a simple correlation of glacial = pluvial can be ruled out.

20.7.5 Changes in the biota

One of the most poignant aspects of the late Quaternary paleogeography of North America was the disappearance of many large mammals from the continent at about the time that the climate was becoming warmer, the ice caps were receding, the plant cover was becoming more luxuriant and prehistoric hunter-gatherers from eastern Siberia were starting to occupy the land. The existence of well-dated mammoth butchery sites in the north of the continent has been used as evidence that the makers of the Clovis spear points hunted these big animals to the point of extinction (Martin and Wright, 1967; Martin, 1984; Martin and Klein, 1984). A contrary view is that they were unable to adjust to the rapid changes in climate following the Last Glacial Maximum, which altered the vegetation and reduced their habitat. The obvious counter to this argument is that there were similar rapid changes in climate and environment during previous glacial-interglacial cycles, all of which they must have survived to be still present in North America during the very late Pleistocene.

More recent work involving the analysis of ancient DNA has shown that mammoths and horses survived in the Alaskan interior until at least 10,500 years ago, or several thousand years after the first arrival of the prehistoric hunters (Haile et al., 2009). Similar studies of ancient DNA from 1,439 well-dated faunal sites in Eurasia showed that climate change alone was adequate to explain the extinction of the Eurasian musk ox and woolly rhinoceros, but a combination of climatic and human impacts caused the extinction of the steppe bison and the wild horse (Lorenzen et al., 2011). The use of ancient DNA to resolve the questions raised by Martin (1984) will likely lead to novel insights and substantial modifications of earlier conclusions. Other approaches are also being pioneered with a view towards establishing a more precise chronology

of climatic changes. For example, Polyak et al. (2012) used the $\delta^{13}\text{C}$ and $\delta^{234}\text{U}$ values in speleothem calcite from Fort Stanton Cave in southern New Mexico as an index of effective precipitation, together with ages obtained from rim pools in the Big Room of Carlsbad Cavern. They discovered that following a moist climatic interlude, a very severe drought was evident in the arid American Southwest from just before 14.5 until around 12.9 ka, coeval with the warm Bølling/Allerød interval evident in the Greenland ice core records from 14.6 to 12.8 ka. They pointed out that the last appearance of sixteen out of thirty-five mammal genera that became extinct between 13.8 and 11.4 ka coincided with this 1,500-year drought in the American Southwest and took place before the arrival of the Clovis hunters in this region. It also occurred before the controverted cometary impact that Firestone et al. (2007) proposed as the reason for the demise of the large late Pleistocene mammals of North America.

An important archive that has been used to reconstruct the late Quaternary ecological and climatic history of the deserts of northern Mexico and the arid south-west of the United States is that provided by packrat middens (Betancourt, 1990a; Betancourt et al., 1990a; Betancourt et al., 1990b; Spaulding, 1990; Van Devender, 1990a; Van Devender, 1990b). In the very arid Chihuahuan Desert, 220 packrat middens and 259 associated AMS ^{14}C ages have revealed a 40 ka history of vegetation change. The lowest parts of the desert remained arid throughout this time, but at higher elevations, the macroflora preserved in the middens show that early Wisconsin climates were somewhat wetter than those in the middle Wisconsin at around 31 ka, with humid climates during full glacial times at about 22 ka. The presence of C_4 perennial grasses points to rainfall in late spring or summer, when temperatures were relatively warm. Van Devender (1990a) concluded that the LGM climate was comparatively mild, with few winter freezes, cool summers and higher-than-present rainfall throughout the desert. This conclusion runs counter to the earlier views of Galloway (1970; 1983) and Brakenridge (1978), who argued for a cold, dry LGM climate. The winter rainfall regime ended sometime after 9–8 ka, with the establishment of the modern climatic regime by about 4,000 years ago. Van Devender (1990a) noted that the response to climate changes varied with the species, so the resulting plant communities were always in a state of flux and never attained equilibrium, a conclusion reached earlier by Davis (1976; 1986) from her studies of postglacial deciduous forest change in North America. A more recent record of Holocene environmental change comes from the $\delta^{13}\text{C}$ composition of two stalagmites from a cave in central Missouri, which show positive excursions at 3.5 and 1.2–0.9 ka that are consistent with more arid climates in the semi-arid Great Plains at those times (Denniston et al., 2007).

20.8 Conclusion

The Cenozoic cooling and desiccation of North America was closely associated with the tectonic movements that caused uplift of the Rockies along the west of the continent, faulting and subsidence in the Basin and Range Province, and progressive

cooling of the Arctic Ocean from 33.5 Ma ago onwards. The cooling culminated in the relatively rapid accumulation of ice across the Laurentian shield around 2.5 Ma ago, resulting in the formation of a major ice cap. The waxing and waning of successive ice caps was modulated by changes in the earth's orbital geometry, with the 23 and 19 ka precessional cycles dominant until 2.6 Ma, the 41 ka obliquity cycle dominant between 2.5 and 0.7 Ma, and the 100 ka orbital eccentricity cycle dominant thereafter. Deposition of loess and wind-blown desert dust was characteristic of cold, dry periglacial environments, with dunes active during glacial times. Lakes in the Great Basin were generally high during times of maximum glaciation, although there were regional differences in the timing of late Pleistocene high lake levels as a result of differences in precipitation source areas. The most recent ages obtained from late Pleistocene high lake levels do not support the once widely accepted hypothesis of a southward displacement of the westerlies during glacial times. In the presently very arid deserts of northern Mexico and the American Southwest, the macrofossil remains in packrat middens indicate relatively mild and moist conditions during the LGM, in opposition to an early view espousing a cold, dry LGM climate in the American Southwest. Both lake shorelines and fossil soils that developed on alluvial fan deposits indicate that the climate was wetter during earlier glacial cycles than it was during the LGM. The precise causes of megafaunal extinctions in North America remain enigmatic, with evidence from ancient DNA and from cave speleothems pointing to an important role played by climate as opposed to simply attributing all extinctions to the arrival of the Clovis hunters.

21

South American deserts

We here [in Peru] have unequivocal evidence that a ridge had been uplifted right across the old bed of a stream . . . and a new channel formed. From that moment, also, the neighbouring plain must have lost its fertilising stream, and become a desert.

Charles Darwin (1809–1882)
The Voyage of H.M.S. Beagle
(3rd ed., 1845)

21.1 Introduction

South America is the only continent that extends from equatorial to high southern latitudes and thus provides a record of continental climatic fluctuations spanning 68 degrees of latitude, from 12°30'N to 56°S. The Atacama Desert is the driest desert on earth. It may also be the oldest (Houston and Hartley, 2003; Hartley et al., 2005). Its history is intimately associated with that of the Andes, which run parallel to the western coastline for 7,000 km and have always exerted a major influence on regional climate. This chapter gives an overview of present-day environments in South America and of the Cenozoic evolution and Quaternary climatic fluctuations of this unique and fascinating continent. The historic floods and droughts linked to El Niño-Southern Oscillation events (for which there is a 500-year annual record from Peru) are covered in [Chapter 23](#) and are therefore not discussed here.

21.2 South American landscapes

Because the climates of South America, including those of the drier regions of the continent, are so strongly influenced by its topography, we begin with a brief account of the major topographic elements of South America. The three dominant elements of the South American landscape are the Andes Cordillera to the west, the plateaux of the Precambrian Shield areas to the east and the lowlands between the Andes



Figure 21.1. Major physiographic regions of South America.

and the plateaux, which include the vast forested alluvial plains of the Amazon, Orinoco and Paraná rivers (Figure 21.1). The Andes are the dominant feature in the South American landscape, with an average elevation of 4,000 m, rising to 6,962 m (Mount Aconcagua), and are high enough to sustain perennial snow and ice for much of their length. They consist of a single set of mountain ridges in the south, two sets of ridges separated by broad upland basins in the centre (Helmens and van der Hammen, 1994) and three sets of ridges in the far north, where the Andes change direction and trend from west to east, parallel to the Caribbean Sea. The hyper-arid Atacama Desert lies in the western rain shadow of the central Andes between latitudes 15°S and 30°S. Its origin is bound up with the tectonic evolution of the Andes. The Patagonian semi-desert lies east of the southern Andes in the rain shadow of the westerlies and is a land of wind and dust, although far less arid than the Atacama. In its absence of trees, it is reminiscent of the arid Nullarbor (‘treeless’) Plain of southern Australia, but it is a great deal windier. The Bolivian Altiplano (‘High Plain’) is an elevated arid plateau, flanked by high mountain ranges which rise to more than 6,000 m, with scattered freshwater and salt lakes, or *salar*s, as well as the remains of much larger lakes (Sylvestre et al., 1999; Sylvestre, 2009). It is 1,200 km long from north to south, 300 km wide from east to west and has an average elevation of about 3,800 m. It lies between the Western and Eastern Cordilleras in the central Andes.

The Amazon Basin supports the largest tropical rainforest in existence but has on occasion been dry enough for sand dunes to develop locally, most probably as source-bordering dunes (Iriondo and Latrubesse, 1994; Latrubesse and Ramonell, 1994; van der Hammen and Hooghiemstra, 2000; Teeuw and Rhodes, 2004). North-east Brazil is prone to severe droughts during El Niño years and has been much drier at intervals during the Quaternary (Auler and Smart, 2001; Wang et al., 2004). The coastal strip of Peru is very arid, but during El Niño years it can receive heavy downpours and is prone to landslides along the Andean foothills.

Vegetation is controlled by temperature, precipitation and evaporation, which are in turn closely linked to relief. On the windward side of the Andes, precipitation increases with elevation, and the same is true of the coastal escarpments. However, temperatures also decrease as elevation increases, with more rapid rates of decrease occurring in the drier regions (Houston and Hartley, 2003). As a consequence, there is a well-marked altitudinal zonation in the vegetation, with mountain forest giving way to grassland and to alpine scrub (Hooghiemstra and Ran, 1994, fig. 2). In the lowlands, tropical rainforest gives way to savanna as rainfall decreases, ceding ultimately to the low shrubs of Patagonia and the sporadic halophytes of the Atacama.

21.3 Present-day climate and causes of aridity

The southern third of the continent (south of about latitude 30°S) falls within the domain of the westerlies, which bring winter rainfall to southern Chile and the western slopes of the southern Andes (García, 1994; Sylvestre, 2009). Patagonia lies in the rain shadow of the westerlies. Katabatic winds blowing down from the Andes accentuate the effects of low rainfall, as does the presence of the cold Malvinas/Falklands Current off the east coast, so this region is sparsely vegetated and exceptionally windy. During earlier, even drier intervals in the Quaternary, Patagonia was a major exporter of dust to Antarctica.

The northern half of the continent is under the influence of the south-east Trade Winds, which bring rain to the east coast of Brazil but blow parallel to the coast on the west, leading to upwelling of cold ocean water and enhanced aridity. Seasonal displacement of the Intertropical Convergence Zone (ITCZ) brings summer rainfall to the region south of the Amazon Basin, which is an area of perennial rainfall (Figure 21.2). The southern boundary of the ITCZ in January at the height of the austral summer extends to 10°S across Amazonia but only to just south of the equator on the east coast and a few degrees north of the equator on the west coast (García, 1994). The Andes therefore lie within the zone of westerly precipitation in the far south and the zone of easterly precipitation controlled by the seasonal migration of the ITCZ in the centre and north (Figure 21.3). As a consequence, snow accumulation will reflect winter westerly precipitation in the south and summer easterly precipitation in the centre and north of the continent. In addition, the Atacama and Peruvian coastal

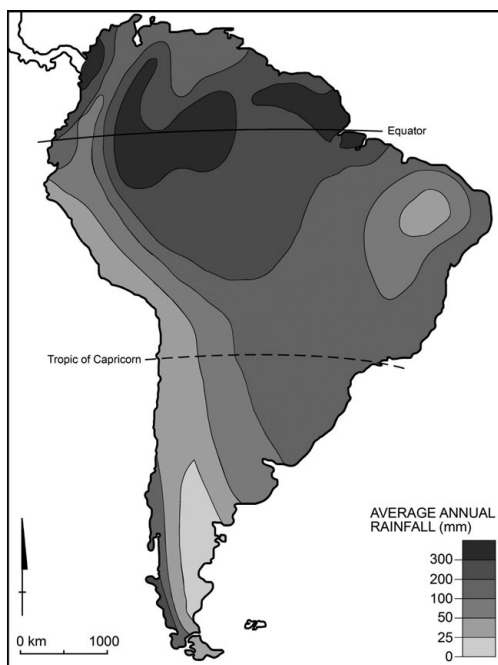
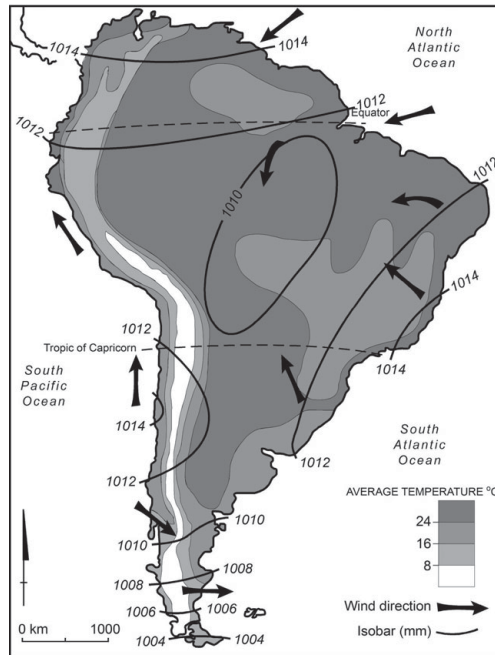


Figure 21.2. Mean annual precipitation, South America. (From *The Times Atlas of Americas*, 2010.)

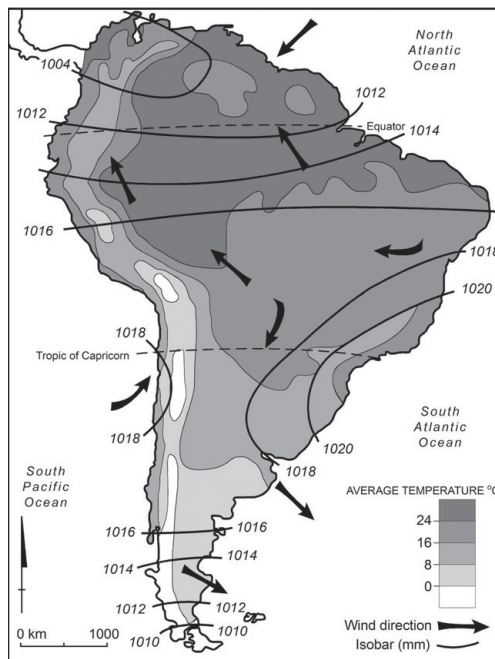
deserts lie within the zone of the eastern rain shadow of the Andes in the centre and north (Figure 21.2). The precise limits of these rain-shadow zones have probably fluctuated during the Quaternary according to how far north the westerly air masses reached in winter and how far south the ITCZ reached in summer. Such changes need not have been in phase.

In addition to topography and seasonal shifts in atmospheric circulation, the ocean currents close to the coast have a significant influence on regional climate. In particular, the cold Humboldt/Peru Current accentuates aridity along the west coast and the cold Malvinas/Falklands Current accentuates aridity off the east coast of Patagonia. In contrast, the warm South Equatorial Current and its southern counterpart, the warm Brazil Current, favour the passage of warm, moist air masses onto the north-east and south-east tropical margins of the continent. In the far north-west, a warm current known locally as the El Niño Current flows from the north, and in years when it reaches the arid coast of northern Peru, it is associated with heavy local rainfall (see Chapter 23).

The Atacama Desert owes its aridity to four main factors. In common with all the other hot tropical deserts of the Southern Hemisphere, it lies beneath the descending southern arm of the Hadley Cell, which brings warm, dry air to that latitude (see Chapter 1). As noted in the previous section, it lies within the western rain shadow



(a)



(b)

Figure 21.3. Surface wind, temperature and pressure patterns, South America (a) during the southern summer (January) and (b) during the southern winter (July). (From *The Times Atlas of Americas*, 2010.)

of the Andes and is flanked offshore by a cold ocean current that further accentuates its aridity. Finally, it lies several thousand kilometres inland with respect to moist air masses from the north-east, which bring heavy rainfall to the eastern flanks of the northern Andes and to the lowlands of Amazonia.

21.4 Cenozoic tectonism, desiccation and cooling

The inception of aridity in the Atacama, Patagonia and the Bolivian Altiplano has its roots in the Mesozoic and Cenozoic tectonic history of South America. South America was once part of the supercontinent of Pangea, or Gondwana, which was in existence for several hundred million years before the initial Jurassic break-up some 180 Ma ago. Gondwana split up in several stages, with the eastern portion (comprising South America and Africa) separating some 180 Ma ago from the western portion (comprising Antarctica, Australia and Greater India) to form the Indian Ocean (Smith et al., 1981; Owen, 1983; Kearey and Vine, 1996; Williams et al., 1998). South America and Africa began to move apart in the Early Cretaceous from about 125 Ma onwards, giving rise to the South Atlantic Ocean (Chapter 3; Figure 3.2). Movement of the Nazca tectonic plate eastwards away from the East Pacific Rise and its subduction beneath the South American Plate (Chapter 3; Figure 3.1) resulted in uplift of the Andes and creation of the deep Peru-Chile trough along the west coast of South America. The existing pattern of ocean circulation around South America was largely established by the end of the Cretaceous some 65 Ma ago. The former flow of the warm equatorial current between North and South America from the Atlantic to the Pacific ended with closure of the Panama Isthmus around 3.5 Ma ago. The Antarctic circumpolar current came into being as a result of the opening of the Drake Passage between Antarctica and the southern tip of South America by around 25 Ma. Unlike Australia and Greater India, which moved rapidly away from Antarctica, South America moved only very slowly northwards, amounting to about ten degrees of latitude. As a result, South America has been within the global circulation system of the Hadley Cell in the north and the westerlies in the south for at least 90 Ma (Houston and Hartley, 2003).

The uplift history of the Andes is complex and was not uniform from north to south. Uplift took place in a series of discrete stages separated by intervals of tectonic stability and sustained erosion (Garzzone et al., 2008). A variety of different methods have been used to determine the degree of uplift and when such uplift occurred. These methods include geomorphology (Coltorti and Ollier, 2000; Montgomery et al., 2001; Hoke et al., 2004), sediment analysis (Hoorn et al., 1995; Hartley et al., 2005; Uba et al., 2007), pollen analysis (Helmens and van der Hammen, 1994; Hooghiemstra and Ran, 1994; Hooghiemstra, 1995; Markgraf et al., 1995; Wijninga et al., 2003), analysis of fossil soils (Rech et al., 2006), stable isotope geochemistry (Ghosh et al., 2006b; Garzzone et al., 2008; Poulsen et al., 2010) and thermochronology, backed up by traditional

dating techniques such as magnetostratigraphy, fission track, zircon uranium-lead and potassium-argon dating (van der Hammen and Hooghiemstra, 1995; Uba et al., 2007). Given the many different approaches used and the often unproven assumptions on which some of them are based, it is no surprise that a detailed consensus on the uplift history of the Andes is yet to be achieved, although the broad outlines now seem reasonably clear. Because aridity in the Atacama is in part related to its location in the western rain shadow of the Andes, some workers have used the history of aridity in the Atacama as an indirect measure of Andean uplift, while others have relied on more direct evidence of aridity (Dunai et al., 2005). We will now consider the results of some of these studies.

Hartley et al. (2005) postulated that the present-day location of the Atacama Desert within the dry subtropical belt was the dominant cause of its aridity. They went on to argue that because there had been very little latitudinal displacement of this region since the late Jurassic 150 Ma ago, aridity must have prevailed in the Atacama for the past 150 million years. A contributing factor was the presence offshore of a cold upwelling current, from at least the early Cenozoic onwards. They concluded that the Atacama was the oldest desert in existence. Houston and Hartley (2003) considered the more specific question of when the central Atacama became hyper-arid. They examined the relationship between elevation and precipitation on the eastern and western slopes of the Andes and concluded that hyper-aridity developed progressively with uplift of the Andes, especially after the Andes attained elevations of 1–2 km, which led to a significant rain-shadow effect. In addition, intensification of the cold Peruvian/Humboldt Current between 15 and 10 Ma ago would have led to enhanced desiccation inland between latitudes 30°S and 15°S.

Miocene tectonic uplift in the north-east Andes caused the diversion of drainage directions from north to east, with the Amazon and Orinoco flowing into the Atlantic by the late Miocene (Hoorn et al., 1995). In the Ecuadorian Andes, a major planation surface graded to sea level developed at the end of the Lower Pliocene and was later uplifted to an elevation of 3,500–4,000 m, becoming deeply incised during the Middle and Late Pleistocene (Coltorti and Ollier, 2000). Using the clumped isotopic values in fossil soil carbonates (see Chapter 7), Ghosh et al. (2006b) estimated that the Bolivian Altiplano had risen at a mean rate of 1.03 ± 0.12 mm/year between about 10.3 and about 6.7 Ma. They suggested that uplift of the Altiplano amounted to $3,700 \pm 400$ m in that time. Both the rate and the amount are probably overestimates, given that the isotopic lapse rate (i.e., the change in the $\delta^{18}\text{O}$ content of precipitation with increasing elevation) would have been lower before the inception of convective rainfall associated with the uplift (Poulsen et al., 2010). These latter authors used more realistic precipitation isotopic lapse rates for their analysis of sedimentary carbonate from the Bolivian Altiplano and concluded that the late Miocene $\delta^{18}\text{O}$ depletion that they identified in the carbonate record indicated the onset and intensification of convective rainfall once the plateau had attained an elevation of

about 2 km, which triggered orographic rainfall from the South American low-level jet stream.

Garzione et al. (2008) used a combination of oxygen and deuterium isotopic analyses to determine uplift along a line from the Western Cordillera across the Central Andean plateau to the Eastern Cordillera. They concluded that the elevation of the Altiplano was less than 2 km until 10 Ma, while the Eastern and Western Cordilleras were only 2.5 to 3.5 km high until that time. Relatively rapid uplift took place in the late Miocene after 9 Ma. This inference is consistent with the rapid increase in sediment accumulation in the Andean foothills between 7.9 and 6 Ma, which Uba et al. (2007) attributed to monsoon intensification and greater climatic variability in the late Miocene. In this context, we need to bear in mind that while Andean uplift could have altered climate, climate may also have influenced the morphology of the Andes (Montgomery et al., 2001). Hoke et al. (2004) found that along the western slopes of the Altiplano plateau, a relict pattern of trellised drainage had been cut by deeply incised canyons, or *quebrados*, some of which were in existence during the late Miocene and early Pliocene. These canyons were formed by groundwater sapping processes associated with the drying out of the climate and uplift of the plateau, which led to a steepened hydraulic gradient.

Let us now consider the desiccation record. Alpers and Brimhall (1988), Clark et al. (1990) and Sillitoe and McKee (1996) concluded that the supergene oxidation and enrichment of the Chilean porphyry copper deposits required a climate that was less arid than the present hyper-arid climate of northern Chile and southern Peru, with Clark et al. (1990) suggesting more than 100 mm of precipitation. All of these authors inferred that progressive desiccation had taken place since the supergene enrichment ceased, with somewhat different ages of 35–14 and 15–9 Ma suggested for the timing of this regional desiccation. Dunai et al. (2005) measured the ^{21}Ne in clasts collected from surfaces in the Atacama that were deemed sensitive to potential erosion from surface run-off. The surface exposure ages of the sediments showed minimal evidence of any erosion in the last 25 Ma, indicating sustained aridity of the Atacama from late Oligocene-early Miocene times onwards.

Rech et al. (2006) investigated a series of fossil soils in the central Andes in order to determine the timing of uplift and of climatic desiccation. A series of calcic vertisols with gleyed horizons and calcium carbonate root traces were overlain by gypseous soils with pedogenic nitrate at elevations between 2,900 m and 3,400 m in the Calama Basin on the eastern edge of the Atacama. Calcic vertisols with gleyed horizons develop in poorly drained, vegetated flood-plains under a seasonal rainfall regime (see Chapter 15). Rech et al. (2006) suggested that these soils had formed under a mean rainfall of >200 mm/year. The transition to salic gypsisols with pedogenic nitrate took place between 19 and 13 Ma. Such soils form today at elevations below about 2,500 m, where the rainfall does not exceed 5–10 mm/year, but the fossil soils now occur at elevations 400–900 m higher than this, indicating that amount of uplift in

the last 10 Ma. The desiccation evident in the transition from semi-arid to hyper-arid conditions probably arose when the central Andes had attained an elevation of at least 2 km during the middle Miocene, thereby preventing moist air masses from the summer monsoon from reaching the Atacama.

To sum up, the Cenozoic uplift of the Andes took place in stages, with long intervals of stability punctuated by shorter intervals of relatively rapid uplift. Major uplift during the Miocene resulted in the creation of an effective rain shadow on the western margin of the Andes. Desiccation of the Atacama was underway by around 25 Ma, with extreme aridity setting in between 19 and 13 Ma, once the central Andes had reached an elevation of 1,500–2,000 m. It is possible that the Atacama has been relatively dry for the past 150 Ma. This is because it is located in the subtropical zone of dry subsiding air and has undergone very little latitudinal tectonic displacement in that long interval of time.

Determining the onset of Cenozoic cooling in South America is more problematic than simply reconstructing the history of uplift and desiccation. There was a major reorganisation of the plankton ecosystem in the Southern Ocean at the start of the first major Antarctic glaciation in the earliest Oligocene around 33.6 Ma ago (Houben et al., 2013). However, in southern Argentina, isotopic analysis of fossil mammal teeth revealed no change in temperature across the Eocene-Oligocene transition, prompting Kohn et al. (2004) to propose that the widely held concept of cooling at this time (see Chapter 3) might be in error. Plant macrofossils from Patagonia certainly show very high plant diversity in the early Eocene around 50 Ma ago (Wilf et al., 2003). Nevertheless, the fossil pollen from central Colombia and western Venezuela do indicate a decline in plant diversity at the end of the Eocene and in the early Oligocene (Jaramillo et al., 2006). Because plant diversity fluctuates globally with temperature, this decline may indicate cooling, or it may indicate a change in the area available for tropical plants to inhabit, which is a function of both temperature and precipitation. Markgraf et al. (1995, p. 143) state categorically that ‘by the late Miocene (10 million years ago), cooling had set in, and by Pliocene times, climatic cycles had become increasingly marked’. This generalisation is supported by the detailed pollen analysis of van der Hammen and Hooghiemstra (2000) in Amazonia and by that of Helmens and van der Hammen (1994), Hooghiemstra and Ran (1994), van der Hammen and Hooghiemstra (1995) and Wijninga et al. (2003) in the High Plain (Altiplano de Bogotá) of Colombia.

21.5 Quaternary climatic fluctuations

In common with the rest of the desert world, the last 2.5 Ma of climatic history in arid and semi-arid South America were characterised by rapid fluctuations in temperature and precipitation linked to the glacial-interglacial cycles of the Quaternary (see Chapter 3). In the late Pliocene, the climatic cycles were primarily the

23 and 19 ka precessional cycles, with the 41 ka obliquity cycles dominating from around 2.5 to about 0.7 Ma, after which the 100 ka orbital eccentricity cycles became dominant. As a result, the earlier portion of the Quaternary from 2.5 to 0.7 Ma was characterised by high-frequency, low-amplitude climatic fluctuations, while the last 0.7 million years were characterised by low-frequency, high-amplitude climatic fluctuations.

21.5.1 Vegetation history

Deposition within the Bogota Basin has preserved a remarkably complete record of river, lake and swamp sediments, as well as associated fossil pollen and spores, allowing the reconstruction of the longest and most complete record of Quaternary tropical vegetation changes for any place on earth (Helmens and van der Hammen, 1994; Hooghiemstra and Ran, 1994; van der Hammen and Hooghiemstra, 1995; Wijninga et al., 2003). As a very broad generalisation, the glacial maxima were cold and generally dry in the lowland tropical regions of South America, while the interglacial maxima were warm and relatively humid. In the southern Andes and parts of the central Andes, the LGM was cold but humid (Sylvestre, 2009, fig. 1.2). Complicating this interpretation were periodic episodes of rapid erosion and landslides that were linked, directly or indirectly, to the episodic uplift of the Andes. Although many workers consider that even the lowland rainforests of Amazonia were adversely affected by episodes of aridity (Iriondo and Latrubesse, 1994; Latrubesse and Ramonell, 1994), it is not always easy to separate out the net effects of changes in temperature, evaporation and precipitation, and Colinvaux and his co-workers have always stoutly denied that the Amazon plains were ever arid (Colinvaux et al., 1996; Colinvaux et al., 2000; Colinvaux, 2001).

The pollen preserved in lake sediments from Lago Condorito in north-west Patagonia provides a record of millennial-scale vegetation and climate changes for the past 15 ka (Moreno, 2004), with evidence of a cool-temperate, humid climate between 15 and 11 ka, followed by a very warm and dry phase between 11 and 7.6 ka, with later cooling and then renewed aridity between 2.9 and 1.8 ka. Moreno (2004) suggested that the millennial-scale climatic changes evident in north-west Patagonia were associated with changes of similar duration in the tropical Pacific, which were related to corresponding changes in the mid-latitude South Pacific – a working hypothesis for future testing.

21.5.2 Desert dust and loess

Iriondo (1993) has long argued that the presence of dunes in both the Amazon Basin and in the vast tropical plains of the Chaco in Argentina, Paraguay and Bolivia is indicative of previously drier climates, as are the thick loess deposits of the Chaco.

Teeuw and Rhodes (2004) obtained OSL ages for dunes in the savanna region of north-east Amazonia, with the start of eolian activity between 17 and 15 ka and continuing until the present. They were unable to decide whether or not this record of gradual eolian deposition had any climatic significance. Where dunes and loess deposits are closely associated, as in the Chaco (Iriondo, 1993), it is more probable than not that the climate was dry and windy when these eolian sediments were accumulating. During times of low glacial sea level, the source area for dust from Patagonia doubled, and considerable volumes of dust reached Antarctica at those times (Maher et al., 2010). The supply of dust from Patagonia diminished rapidly during postglacial times when meltwater lakes occupied the Patagonian glacial outwash plains at the foot of the glaciated mountains (Sugden et al., 2009).

21.5.3 Glaciations

The southern Andes between latitudes 27°S and 55°S have a record of glacial activity extending from the Late Miocene until the present, making it one of the most complete records on earth, with ages based on magnetostratigraphy and potassium-argon (K-Ar) dating (Rabassa and Clapperton, 1990; Clapperton, 1993). Glacial till has been found between lavas with K-Ar ages of 7.0–4.6 Ma, and younger tills have ages of 4.5, 3.68–3.55, around 3.5, 2.6–2.0, 2.05–1.03, 1.36–1.32 and around 1.2 Ma. This latter Early Pleistocene glaciation seems to have been the most extensive of all the Patagonian glaciations. The Early and Mid-Pleistocene glaciations have been distinguished using relative dating methods such as weathering rind thickness on moraine boulders and the degree of iron and manganese crust formation. Upper Pleistocene glacial advances have ages of around 70, 20–18 and 15–10 ka, with the LGM (the Llanquihue Glaciation) being synchronous with the Wisconsin and Weichsel glaciations of North America and Europe (Rabassa and Clapperton, 1990) and corresponding Marine Isotope Stage 2 (MIS 2) glaciations in Australia and New Zealand (Clapperton, 1990). Snow-line depression during the LGM amounted to roughly 1,000 m. The MIS 4 glacial advances were often more extensive than those of MIS 2, because of lower temperatures and/or greater precipitation during MIS 4 along the southern Andes.

Much of this earlier work was characterised by outstanding field mapping of glacial moraines but suffered from the lack of an adequate chronology of glacial advances and retreats. In the last decade or so, the application of cosmogenic nuclide dating to boulders within glacial moraines has revolutionised research on past glacial activity in South America, as indeed elsewhere. Murray et al. (2012) obtained ^{10}Be exposure ages for boulders on the crests of moraines laid down by former glaciers in the Rio Guanaco Valley of southern Patagonia in latitude 50°S. The LGM ended in this valley by 19.7 ± 1.1 ka and rapid glacier retreat had started by 18.9 ± 0.4 ka, with half of the upper valley ice retreat achieved by 17.0 ± 0.3 ka. Retreat of the Laurentide ice

cap in North America at around 19 ka caused changes in the ocean circulation in the Atlantic, resulting in warming at high latitudes in the Southern Hemisphere, followed by the melting of ice caps and glaciers in Patagonia.

^{10}Be exposure ages obtained from moraines of outlet glaciers in the South Patagonian Ice Field at 51°S showed that they advanced during the Antarctic cold reversal (around 14.6–12.8 ka) and were retreating rapidly by 12.5 ka, consistent with temperature changes in Antarctica and the Southern Ocean at this time (García et al., 2012). Precipitation dominantly came from westerly air masses, which reached much further north during the LGM, when the Subtropical Front was at about 40°S, and fluctuating after that in response to changes in the SubAntarctic Front and the Polar Front (García et al., 2012).

Ramage et al. (2005) assessed different methods of determining the former snow-line or equilibrium line altitude (ELA) in what is today an ice-free part of the tropical Andes of central Peru. They concluded that during the LGM, the ELA had been 220–550 m lower than it is today, pointing to a slight temperature decrease of about $2.5 \pm 1^\circ\text{C}$. The ELA lowering estimated for the LGM was similar to that of the most extensive glaciations in these valleys with ages in excess of 65 ka. The obvious inference from this is that the relative influence of temperature and precipitation on snow accumulation was not the same at these two times. These results from the Peruvian Andes are in strong contrast to the results obtained by Stansell et al. (2006) for ELA lowering during the LGM in the Venezuelan Andes north of the equator, which showed that the ELA levels were about 1,420 to 850 m lower than present, indicating that temperatures were possibly $8.8 \pm 2^\circ\text{C}$ cooler than at present. A possible reason for this difference is that the Peruvian Andes are much drier than the northern Andes, limiting the potential accumulation of snow and ice.

Using cosmogenic ^3He , Bromley et al. (2011) have shown that there was a readvance or prolonged standstill of glaciers in the arid Andes of south-west Peru in the very late Pleistocene, with moraines located about midway between the LGM and present-day limits dated to 12.8 ± 0.7 ka. Two sets of Holocene moraines in the Peruvian Andes have yielded high-precision cosmogenic ^{10}Be surface exposure ages (Licciardi et al., 2009), with the older moraines dated between 10 and 8 ka and the younger ones being coeval with the latter half of the Little Ice Age of Europe, which is dated between about 1300 and 1860 AD (Lamb, 1977; Grove, 1988). Jean Grove (2004) has shown in abundant detail that there were multiple glacial advances in both hemispheres during the Holocene, so it is likely that future work will provide a more complex pattern of glacial advances and retreats in the Peruvian Andes.

Several ice caps in the central Andes have yielded a useful record of late Pleistocene and Holocene fluctuations in temperature and precipitation, notably the Illimani and Sajama ice caps in Bolivia and the Quelccaya and Huascarán ice caps in Peru (Thompson et al., 1995; Thompson et al., 1998). Analysis of the isotopic composition of the ice ($\delta^{18}\text{O}$ and δD) indicates full glacial conditions at 20–18 ka and an inferred drop in temperature of 8–12°C followed by early Holocene warming. One

problem involved with obtaining quantitative estimates of changes in precipitation and temperature from analysis of the $\delta^{18}\text{O}$ composition of ice cores, alluded to in [Section 21.4](#) when discussing carbonates in fossil soils in the Andes, involves the choice of $\delta^{18}\text{O}$ precipitation lapse rate. This is because elevation is only one of a group of factors controlling the lapse rate, some of the others being distance travelled by the air mass and the intensity of convectional downpours during the passage of moist air across Amazonia. These difficulties apply equally to the paleoclimatic analysis of speleothems collected from limestone caves in South America (Cruz et al., 2009).

Thompson et al. (2013) have recently obtained an annual record of climatic fluctuations for the last 1,800 years from the Quelccaya ice cap. The $\delta^{18}\text{O}$ oxygen isotopic values showed an increase in snow accumulation rates and a decrease in temperature during the latter half of the Little Ice Age, from 1681 to 1880 AD. High concentrations of nitrate and ammonium, and ^{18}O depleted isotopes in the ice cores, coincided with decreased percentages of Ti in the Cariaco Basin marine record, and vice versa. This indicates that when run-off and precipitation along the north-east coast of tropical South America were reduced, conditions were wetter further south over the Amazon Basin, and conversely.

21.5.4 Lake fluctuations

A number of lakes located at high elevations in the central Andes and semi-arid Bolivian Altiplano provide evidence of late Quaternary fluctuations in temperature and precipitation (Sylvestre, 2009). Many of these lakes are now shallow and saline salt pans known as *salar*s. Lake Titicaca was deep and fresh between 25 and 15 ka (Baker et al., 2001). In the southern Bolivian Altiplano, the *salar* of Coipasa was a shallow salt lake between 24.6 and 20.9 ka, when conditions were more humid than today (Sylvestre et al., 1998; Sylvestre, 2002). The *salar* of Uyuni was also less arid between 26.1 and 14.9 ka. The lake chronology for the Uyuni-Coipasa Basin was based on both ^{14}C and $^{230}\text{Th}/^{234}\text{U}$ ages (Sylvestre et al., 1999). One difficulty that Sylvestre and her colleagues noted during this work was the varying reliability of radiocarbon ages. For example, the lakes began to rise a little before 16,000 radiocarbon years ago (^{14}C yr BP) and reached maximum levels between 13,000 and 12,000 ^{14}C yr BP. Following a dry spell, the lake rose again to a lower level between about 9,500 and 8,500 ^{14}C yr BP. There was good agreement between the ^{14}C and $^{230}\text{Th}/^{234}\text{U}$ ages for the first and highest lake phase but a lack of accord for the second lake phase (Sylvestre et al., 1999). They concluded that this discrepancy was probably caused by a delayed response of the groundwater-table during the early Holocene dry phase and used a correction of about 2,000 ^{14}C years for this reservoir effect.

Other workers in the southern Andes and Atacama have faced the same problem of very large reservoir effects. Geyh et al. (1999) investigated the timing of the late glacial/early Holocene humid phase along a high-altitude transect between 18°S and



Figure 21.4. San Pedro de Atacama and Licancabur, Chile. (Photo: Mike Smith.)



Figure 21.5. Rio Loa, central Atacama, Chile. (Photo: Mike Smith.)



Figure 21.6. Geoglyphs, northern Atacama, Chile. (Photo: Mike Smith.)

28°S in the Atacama Desert of northern Chile. In order to establish the magnitude of the radiocarbon reservoir effect, they obtained ^{14}C ages for different types of sample, including non-aquatic, carbon-rich sediments. They found that the reservoir effect ranged from $-1,200$ to $-10,700$ ^{14}C years. Any uncorrected ages would thus seem much older than their true age. On the basis of their reservoir corrections, Geyh et al. (1999) concluded that the humid phase began between 13,000 and 12,000 ^{14}C yr BP, while maximum lake levels were reached between 10,800 and 9,200 ^{14}C yr BP. These results differ from those of Sylvestre et al. (1999) for the timing of the wettest phase. A probable reason for these differences is that the lakes and swamps in the two regions studied were under the influence of air masses that originated from quite separate sources and were active at quite different times, a conclusion also reached by Sylvestre (2009) in her review of LGM climates in South America and by Thompson et al. (2013) in their analysis of Andean ice cores.

21.6 Conclusion

The climate of South America is dominated by the easterly summer monsoon in the tropical north of the continent and by winter westerly air masses in the south of the continent. As a result, the lands to the west of the central and northern Andes are in the rain shadow of these very high mountains, while Patagonia in the far south lies in the rain shadow east of the Andes. The aridity of the Atacama Desert (Figures 21.4, 21.5 and 21.6) results from four independent factors: its location in the zone of dry

subsiding air along the southern limb of the Hadley Cell; the presence offshore of a cold ocean current and cold upwelling ocean water; its distance from the easterly air masses that bring rain to Amazonia; and its location in the rain shadow of the Andes. The Atacama was already arid more than 30 million years ago and may have been dry for some 150 million years, making it the oldest extant desert on earth. The Neogene uplift of the Andes occurred in stages, with prolonged intervals of stability punctuated by phases of relatively rapid uplift. Uplift accelerated in the mid- to late Miocene, causing major changes in erosion, sedimentation, climate and vegetation. The late Pliocene and Quaternary climates were marked by rapid fluctuations in temperature and precipitation and multiple glacial advances and retreats in the high Andes and Patagonian Andes. Glacial maxima were broadly synchronous in both hemispheres. Lake fluctuations have been hard to date precisely because of large reservoir effects but appear to show cold and relatively dry conditions in much of the tropical north during the Last Glacial Maximum, in contrast to the relatively moist but still very cold conditions in the centre and south.

22

Australian deserts

The human intellect cannot grasp the full range of causes that lie behind any phenomena. But the need to discover causes is deeply ingrained in the spirit of man.

Leo Tolstoy (1828–1910)

War and Peace (1868–1869)

(Trans. Anthony Briggs, 2005, p. 1096)

22.1 Introduction

We saw in earlier chapters that the reconstruction of Cenozoic climatic changes in deserts is based on evidence from land and sea, including desert dunes, wind-blown dust, river and lake sediments, glacial moraines, plant and animal fossils, isotope geochemistry, speleothems, soils and prehistoric archaeology. Chapters 7 to 17 considered the scope and limitations of each type of proxy evidence that has been used to reconstruct desert environmental history. As Williams (2009, p. 799) has noted: ‘The length of the climatic record attainable from the various independent lines of evidence and the time resolution possible with each of them vary widely, and cover a time range from 10^{-2} to 10^7 years’. Given the many gaps in the record and the lack of accurate chronologies, our present understanding of past events is perforce provisional. Bearing in mind these cautions, we now review the Cenozoic history of the Australian deserts, starting with present-day environments. Historic floods and droughts documented by instrumental observations and historical archives are discussed in Chapter 23.

22.2 Present-day climate and causes of aridity

After Antarctica, Australia is the driest continent on earth, but it lacks the extreme aridity of the Atacama, Sahara or Gobi deserts. No part of Australia receives less

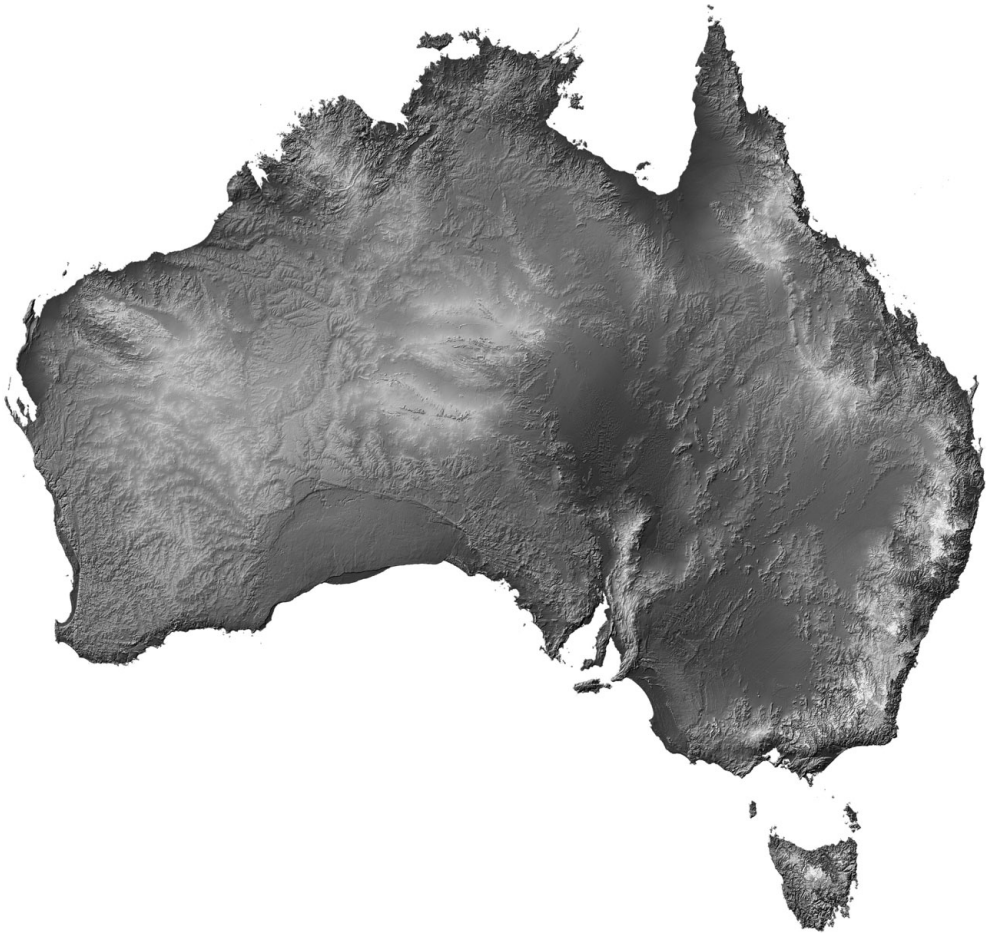


Figure 22.1. Digital elevation model of Australia. (From Byrne et al., 2008.)

than 100 mm of rain a year, which is the value that is used to denote the northern limit of the Sahara. The main factor causing aridity in Australia is its location astride the Tropic of Capricorn, at the southern end of the Hadley Cell (Diaz and Bradley, 2004) in a zone of dry descending air (Chapter 2). Other factors include location in the rain shadow of the Eastern Highlands (Figures 22.1 and 22.2), distance inland and the presence of a cold ocean current off the west coast of Australia. Most of Australia is low and flat, so orographic rainfall is restricted to the Eastern Highlands and to the Flinders and Mount Lofty ranges in South Australia (Figure 22.2). Aridity is greatly accentuated inland with increasing distance from sources of moist maritime air, and rainfall decreases rapidly away from the coastline (Figure 22.3). About 37 per cent of the land receives less than 250 mm/year, 47 per cent less than 375 mm/year and 68 per cent less than 500 mm/year (Gentilli, 1977). The driest region is around Lake Eyre (Figure 22.3; Gentilli, 1977). In the arid interior, rainfall is erratic and

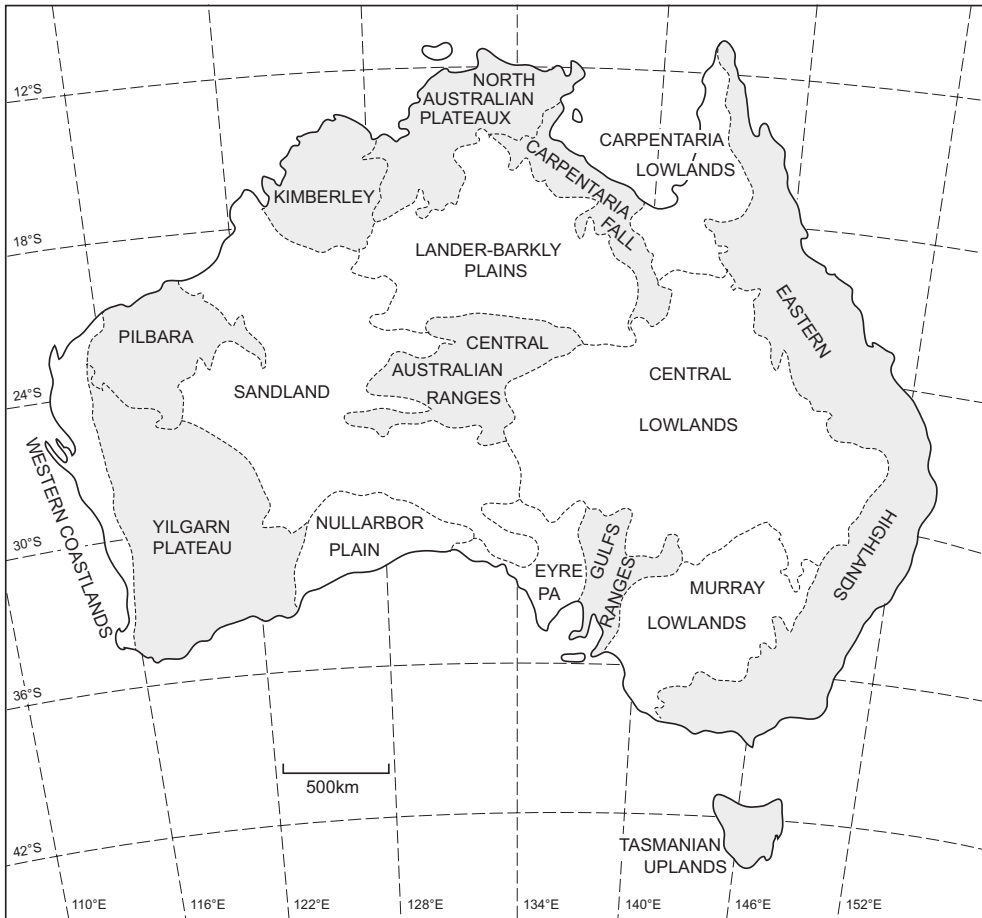


Figure 22.2. Major regions of Australia. (Simplified from Jennings and Mabbutt, 1986, fig. 3.1.)

can occur in both winter and summer. Rainfall variability in the eastern half of the continent is strongly influenced by El Niño-Southern Oscillation events (Allan, 1985; Whetton et al., 1990; Nicholls, 1992; Simpson et al., 1993a; Simpson et al., 1993b; Whetton and Rutherford, 1994; Allan et al., 1996; Kotwicki and Allan, 1998; Cai et al., 2001; Cane, 2005), as well as the North Pacific Oscillation/Pacific Decadal Oscillation (Salinger et al., 2001; Mantua and Hare, 2002; Pierce, 2002), all of which are discussed in Chapter 23.

Australia may be grouped into three broad climatic regions: the seasonally wet tropics, the arid interior and semi-arid north-west, and the temperate southern zone with mainly winter precipitation (Figure 22.3). The seasonally wet tropics occupy the northern one-fifth of the continent and, like the seasonally wet tropics elsewhere, such as the southern margin of the Sahara, have two pronounced seasons: a hot, wet

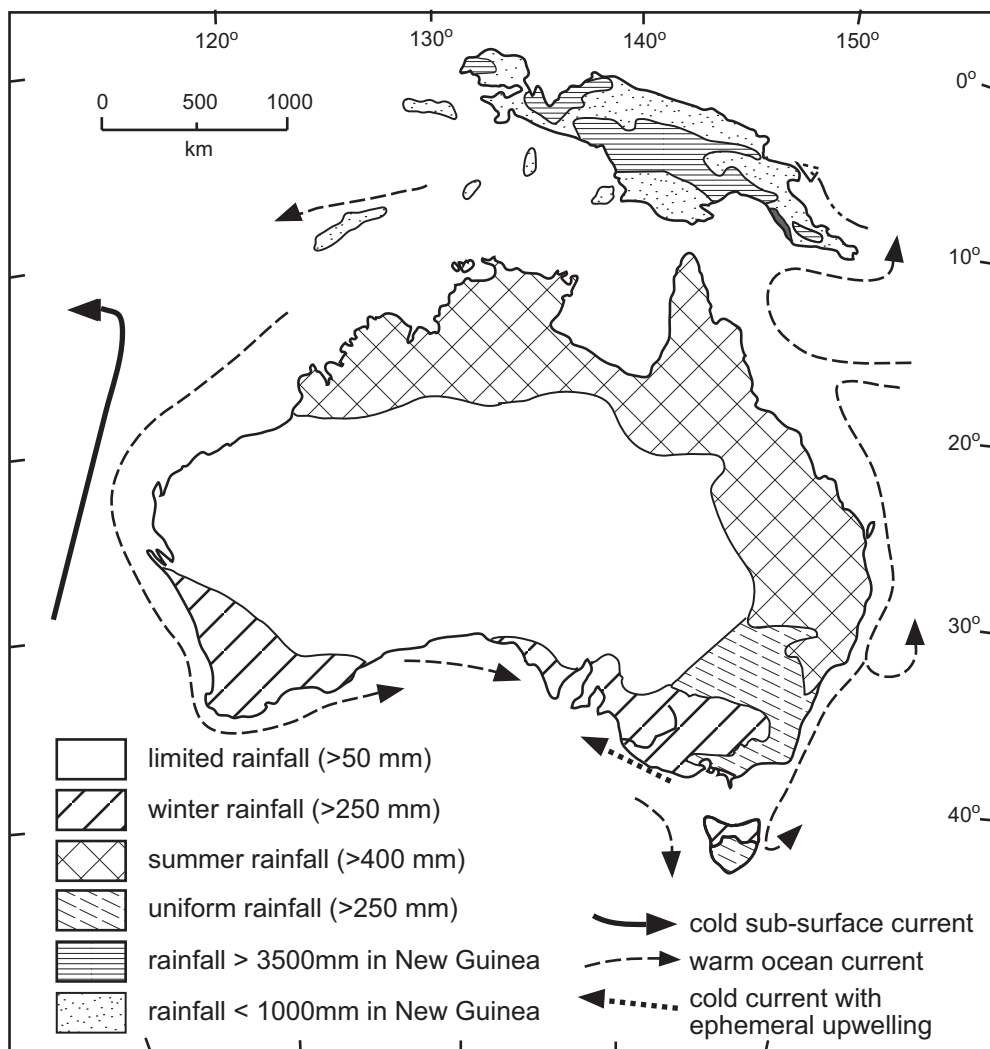


Figure 22.3. Present-day precipitation zones of Australia and surrounding region and major ocean currents. (After Williams et al., 2009b.)

summer and a warm, dry winter. The Intertropical Convergence Zone (ITCZ) lies just north of Australia in January and moves south over tropical Australia in February, which is the peak of the northern Australian wet season (Chapter 2; Figure 2.2). In the far north of Australia, the wet season lasts from November to April and the dry season lasts from May to October, when the land is effectively a desert. In north-east Australia, the summer rains are associated with the southward passage of the ITCZ and with humid easterlies (Gentilli, 1971; Gentilli, 1977; Linacre and Hobbs, 1977; Sturman and Tapper, 1996; Wang, 2006). Rainfall is increased by tropical summer

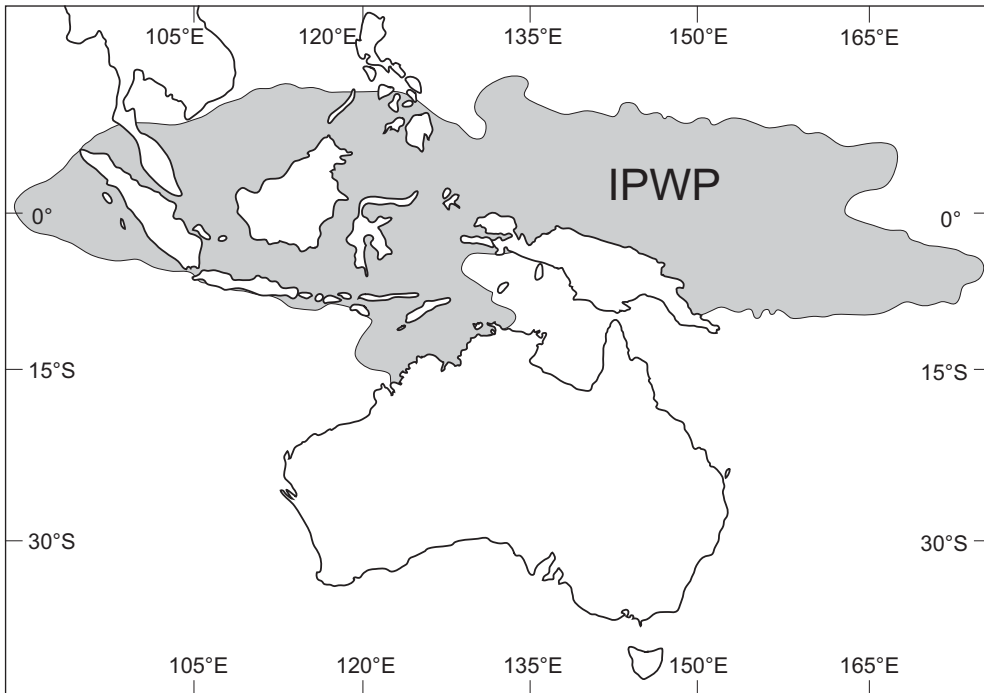


Figure 22.4. The Indo-Pacific Warm Pool bounded by the 28°C isotherm. (After Williams et al., 2009b.)

cyclones and by convective disturbances at the start and close of the wet season (Linacre and Hobbs, 1977). The minimum sea surface temperature (SST) for tropical cyclone genesis is about 26°C (see Figure 22.4), and tropical cyclones can contribute up to 50 per cent of the annual rainfall in the far north (Gentili, 1971). The West Pacific Warm Pool, shown in Figure 22.4, and the poleward flowing warm Leeuwin Current (Figure 22.3) both came into being some 30 million years ago as a result of Australia's northward drift, which is discussed in Section 22.4.

Occasional low-pressure troughs extending from the north-west and north into southern Australia can bring tropical moisture beyond the monsoon zone during summer. The arid centre has very hot, dry summers and cold, dry winters, while the temperate south receives most of its rain in winter, when the westerly air masses extend furthest to the north. A small area in the south-east is also under the influence of the south-east Trade Winds in summer, and receives as much rain in summer as it does in winter.

In order to answer the questions of when, how and why Australia became arid, we need to know when it attained its present latitudinal position, when the Eastern Highlands first attained their present elevation and when the cold West Australian Current was first established.



Figure 22.5. Flinders Ranges, South Australia. (Photo: Frances Williams.)

22.3 Australian desert landscapes

As with other deserts, it is convenient to group the Australian desert landscapes into erosional and depositional landforms (Mabbutt, 1977; Williams, 1984d; Mabbutt, 1988). The erosional components include the dissected rocky plateaux of the Yilgarn and Pilbara regions in the west, the Musgrave, Arunta and MacDonnell ranges in the centre, and the Flinders Ranges in the south (Figures 22.5 to 22.7). The rocks comprising these uplands range in age from Archaean and Proterozoic to Palaeozoic, and they are in general highly resistant to erosion under the present arid climate. Some of the rocks in the Yilgarn have an age of 4.2×10^9 years, making them among the oldest rocks on earth. As a result of this observed resistance to present-day erosional processes, some workers have concluded that parts of the Australian landscape have persisted virtually unmodified since the early Mesozoic or even earlier (Twidale and Campbell, 1991; Williams, 1991; Twidale, 1998; Twidale, 2000). This hypothesis can now be tested using measured rates of erosion.

Heimsath et al. (2010) measured the cosmogenic ^{10}Be produced in situ across a range of rock types and climatic zones in arid Australia. They obtained mean erosion rates of about 1.5 m Ma^{-1} on rocky, weathering-limited slopes in arid central Australia and rates of $2\text{--}11 \text{ m Ma}^{-1}$ for blocky quartzite slopes in the semi-arid Flinders Ranges



Figure 22.6. Kata Tjuta ('The Olgas'), central Australia. (Photo: Frances Williams.)



Figure 22.7. 'Devil's Marbles' granite tor, Northern Territory, Australia.

of South Australia. These compare with mean rates of 35 m Ma^{-1} from soil-mantled, transport-limited spurs in the humid south-east of the continent.

Directly measured annual rates of slopewash on undisturbed granite and sandstone slopes in the seasonally wet tropics of northern Australia were $54 \pm 40 \text{ m}^3 \text{ km}^{-2}$ on granite and $56 \pm 30 \text{ m}^3 \text{ km}^{-2}$ on sandstone (Williams, 1973a). These rates would amount to an average rate of about $55 \pm 35 \text{ m Ma}^{-1}$ for both rock types if extrapolated over longer time scales, but the climate and plant cover would of course have varied during that time, as would rates of slope erosion. Slope lowering by slopewash was on average five times more rapid than that effected by soil creep, and slope lowering was twice as fast on the colluvial-alluvial sandstone foot-slopes than it was on the rocky, weathering-limited hill slopes (Williams, 1973a; Williams, 1976b).

The key lesson to be drawn from studies of erosion rates is that even resistant uplands in the arid interior are unlikely to retain their original form, unless they have been buried beneath thick layers of sedimentary rocks and have only recently been exhumed (Belton et al., 2004). A case in point is the recent emergence of the Lower Proterozoic hills and valleys from beneath the Upper Proterozoic sandstones of the west Arnhem Land escarpment (Williams, 1991).

Other widespread erosional landforms include the extensive plateaux developed on horizontal Mesozoic sandstones and siltstones, many of which have resistant duricrust cappings of silcrete or ferricrete (see Chapter 15). Because the surfaces of these plateaux are covered in a thin layer of silcrete or ferricrete pebbles, they are known locally as stony tablelands, and they are the geomorphic equivalent of the *hamada* of the Sahara and Near East and the *gobi* plains of Mongolia.

The depositional landforms are a great deal younger and consist of dunes and sand plains, lakes and playas, and drainage systems ranging from long-defunct to ephemeral or seasonal. Associated with the lakes or former lakes are crescent-shaped clay dunes, or lunettes, and source-bordering dunes are a feature of many seasonal river channels.

22.4 Mesozoic and Cenozoic tectonism and volcanism

The desiccation of the Australian continent is closely related to its tectonic history (Veevers, 1984; Williams, 1984d; Veevers, 2000a; Veevers, 2000b; McGowran et al., 2004; Fujioka and Chappell, 2010; Quigley et al., 2010a; Blewett, 2012). Until about 100 million years ago, in the early Cretaceous, Australia was largely submerged beneath a warm, shallow ocean, and it mainly consisted of three large islands in the west, north and east. The western island included the Archaean and Precambrian shield areas of western Australia. The rocks forming the northern island and a few central uplands ranged in age from Proterozoic to Palaeozoic, and the eastern island consisted primarily of Palaeozoic rocks. The hills and valleys of these uplands have

been subject to erosion for many hundreds of millions of years, in contrast to the much younger landscapes to the south and east. For example, the Cretaceous and Cenozoic rocks in the Lake Eyre Basin have a denudational history of only a few tens of millions of years.

Until about 130 Ma ago, Australia, Antarctica and Greater India were joined together as part of eastern Gondwanaland, a supercontinent that also once comprised Africa and South America (Veevers, 2000a; Veevers, 2000b). Towards the start of the Cretaceous (around 130 Ma), a rift formed between Australia-Antarctica and Greater India, and the Indian Ocean began to develop. Late Carboniferous to Jurassic uparching of the eventual rift margins preceded this rifting, leading to drainage reversal. Rifting was accompanied by volcanic activity in the Perth and Carnarvon basins, with an incursion of shallow seas, followed by deeper waters as the rift margins collapsed during the Late Cretaceous. The sea floor spreading rate began to increase around 80 Ma ago, reaching up to 17.5 cm/year, and it remained rapid until the late Palaeocene, around 53 Ma ago (Johnson et al., 1976; Veevers, 2000a; Veevers, 2000b).

A rift also began to form during the Palaeocene between Antarctica and Australia, which at first spread apart quite slowly (about 4 mm/year) (Veevers, 2000a). Slow northward drift of Australia continued until around 45 Ma ago, when the rate increased by an order of magnitude to 6–7 cm/year (Veevers, 2000a; Veevers, 2000b; McGowran et al., 2004). Late Eocene marine limestones were laid down in the Eucla Basin and onto the Precambrian shield (Veevers, 1984; Veevers, 2000a; Clarke et al., 2003). The Southern Ocean was developing progressively, with the opening of the Tasman Gateway at around 34 Ma and of the Drake Passage between Antarctica and South America around 30 Ma ago (Doake, 1977; Livermore et al., 2005). Antarctic glaciation was well underway by around 34 Ma (McGowran et al., 2004), as a result of the change in ocean currents around Antarctica (Figure 22.8).

At about the same time that rifting was taking place along the southern continental margin of Australia, rifting was also occurring along the eastern edge of the continent, culminating in the separation of the Campbell Plateau, New Zealand and the Lord Howe Rise from Australia, as well as the formation of the Tasman Sea (Jones and Roots, 1974; Roots, 1975; Veevers, 1984; Veevers, 2000a; Veevers, 2000b).

After fully separating from Antarctica around 45 Ma ago, Australia has moved north into dry, subtropical latitudes at a mean rate of 5–6 cm/year, resulting in the birth of the Southern Ocean between Australia and Antarctica. The opening of the Drake Passage led to the establishment of the circum-Antarctic ocean current, which is driven by the prevailing westerly winds. As a result, Antarctica became thermally isolated from warmer ocean waters to the north, and rapid cooling ensued. In the Southern Ocean, the changing isotopic composition of both planktonic and benthic foraminifera indicates major cooling of deep ocean water as well as surface water (Shackleton and Kennett, 1975; McGowran et al., 2004). Cumulative ice build-up in

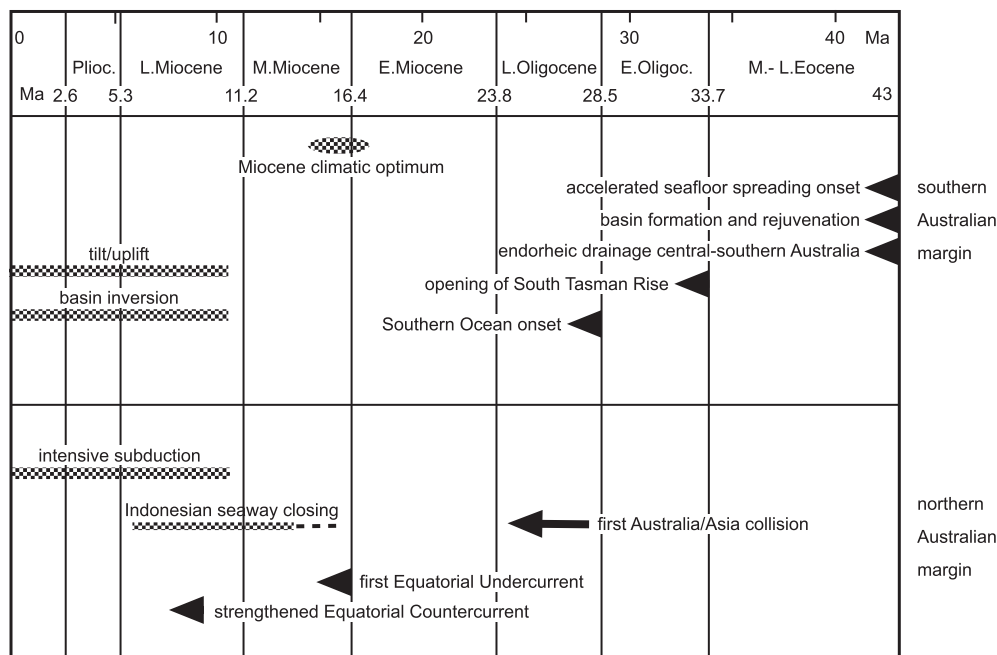


Figure 22.8. Cenozoic environments of southern Australia. (From Williams, 2009a, after McGowran et al., 2004, fig. 9.)

Antarctica saw the creation of mountain glaciers, followed by the growth of a major ice cap, first in East Antarctica 33–34 Ma ago and later in West Antarctica.

As the Australian plate moved northwards, there was intermittent volcanic activity and spasmodic uplift of the Eastern Highlands. Initial uplift of these highlands may have already begun during the late Cretaceous. Uplift showed considerable lateral variation (van der Beek et al., 2001). In the upper Macquarie Basin, in eastern New South Wales, uplift reached a peak during the Middle–Late Miocene, with river incision after that due to climatic and not tectonic causes (Tomkins and Hesse, 2004). Wellman and McDougall (1974a; 1974b) obtained several hundred potassium-argon ages for central volcanoes and valley-fill lavas along the Eastern Highlands, and they found a linear relation between age and latitude, with lavas becoming progressively younger to the south. These ages indicate a northward movement of the Australian plate across one or more stationary hot spots at a mean rate of about 6–7 cm/year, comparable to the rate deduced from sea floor spreading and paleomagnetic data.

Several important conclusions may be drawn from this brief sketch of Mesozoic and Cenozoic tectonic events. The *topographically* controlled pattern of precipitation in eastern Australia is unlikely to have changed very much during the past 30 million years or so. Likewise, the initiation of the mainly westward flowing drainage in eastern Australia extends well back into the Cenozoic. In western Australia, the

Table 22.1. *Cenozoic vegetation and climate in Australia. (Based on data in Alley and Beecroft, 1993; Alley et al., 1996; Alley, 1998; and Martin, 2006.)*

Palaeocene-early Eocene	Warm and humid; mostly warm to cool temperate rainforest; centre seasonal rainfall; north-west dry.
Mid-late Eocene	In centre sclerophyll vegetation on slopes and ridges; rainforest restricted to damp valley floors
Latest Eocene-earliest Oligocene	Abrupt ocean cooling; decrease in megathermal angiosperm diversity across the continent
Early-mid-Miocene	Warm and humid; forest types highly diverse; woodland and Casuarinaceae forests more common; centre warm with high seasonal evaporation; rivers dry out in west; first major onset of aridity
Late Miocene	Cool and dry; rainforests shrink; increase in Eucalyptus/Casuarinaceae sclerophyll forests in central south-east; regular burning of eucalypt forests; rainforests persist along east coast and in highlands; centre more arid with dry woodland and chenopod shrubland
Pliocene	Grasslands develop in central north; rainforests shrink further
Early Pleistocene	Modern climatic regime but wetter; glacial-interglacial cycles with open shrubland/grasslands/herbfields during arid glacials alternating with wooded vegetation during warm wet interglacials
Mid-Pleistocene	Change to drier climate at ~0.5 Ma; still wetter than today.
Last interglacial	Warm and wet; the last major wet interval
LGM	Cold and dry
Holocene	Brief warm, wet intervals

original drainage network that originated between Early Cretaceous and Late Eocene times (130–40 Ma) became progressively disrupted, so that by middle Miocene times, these rivers had ceased to flow and are evident today as a linear series of salt pans (Bunting et al., 1974; Van de Graaff et al., 1977; Salama, 1997; Zheng et al., 1998; Zheng et al., 2002). But the most overwhelming outcome was the movement of much of Australia during the last 30 million years into tropical latitudes dominated by anticyclonic conditions and dry, subsiding air. By about 100 Ma, more than 60 per cent of northern and western Australia was located between 60°S and 40°S. By 80 Ma, only the southern margins of the continent lay south of 60°S, and by about 50 Ma, virtually all of Australia was located between latitudes 60°S and 30°S.

22.5 Cenozoic vegetation history and progressive desiccation

The preservation of fossil spores and pollen in sediments ranging back to the early Cenozoic has made it possible to reconstruct a reasonably comprehensive history of changes in the Cenozoic vegetation of both Antarctica (Truswell and Macphail, 2009) and Australia (Alley and Beecroft, 1993; Alley and Lindsay, 1995; Markgraf et al., 1995; Alley et al., 1996; Alley, 1998; Alley et al., 1999; Martin, 2006). Table 22.1

provides a summary of the Cenozoic vegetation history of Australia and shows that desiccation occurred in a series of steps, effectively starting in the mid-Miocene. Combined with geochemical and sediment analysis, pollen studies in the Lake Eyre Basin have revealed a complex history of deep weathering, formation of ferricrete and silcrete, paleochannel sedimentation and changes in plant cover (Alley et al., 1996; Alley, 1998; Alley et al., 1999). The earliest phase of deep weathering took place before channel alluviation, perhaps in the early Mesozoic. Temperate rainforests flourished along what are now the southern margins of Australia during the Palaeocene, and they gave way to open woodland during the colder, drier Oligocene and Miocene. In central Queensland, there was intense chemical weathering under a presumed warm and humid climate from late Oligocene to middle Miocene (Li and Vasconcelos, 2002). However, clay minerals in sites away from original sources are not always a good guide to climate. John et al. (2006) attributed the high proportion of middle Miocene kaolinite in three marine sediment cores off north-east Queensland to the fluvial reworking of early Miocene lake deposits rich in kaolinite. They concluded that the middle Miocene was a time of global cooling, evident in low glacio-eustatic sea levels. In the central south of the continent, shallow alkaline lakes occupied some of the valley bottoms. Pliocene and Pleistocene desiccation saw the proliferation of open woodland and chenopod shrubland. In a novel approach, Byrne et al. (2008) used evidence from phylogenetics and phylogeography and concluded that the first major indications of aridity date back to the mid-Miocene around 15 Ma ago – a conclusion in accord with the pollen evidence compiled by Martin (2006). The work of Byrne and her colleagues, based on molecular phylogenies, demonstrated that certain taxa show patterns of recent expansion and migration throughout the arid zone, while others seem to have survived in many local refugia during cold, dry glacial times (Byrne, 2008a; Byrne, 2008b; Byrne et al., 2008).

22.6 Quaternary environmental fluctuations

22.6.1 Changes in plant cover

Lynch's Crater in the seasonally wet tropics of north-east Queensland has the longest pollen record of any site in Australia and has provided important information about past climatic fluctuations as well as possible human impacts on the biota (Kershaw, 1976; Kershaw, 1994; Kershaw, 1995; Kershaw et al., 2003a; Kershaw et al., 2003b; Turney et al., 2004; Kershaw et al., 2007). Although it is at present outside the confines of the dry subhumid tropics, as defined in Chapter 1, during cold, dry glacial intervals, precipitation in this locality was reduced to a fraction of that prevailing today (Kershaw and Nanson, 1993). During around 40–30 ka, the moist *Araucarian* rainforests were progressively replaced by open sclerophyll woodlands dominated by *Casuarina* and *Eucalyptus*, reflecting both a drier climate and the impact of burning (Kershaw et al.,

2003a; Kershaw et al., 2003b; Kershaw et al., 2007). The climate remained cool and dry until 19 ka, with open woodlands dominant. The peat humification record from this site, also reflected in changing proportions of Cyperaceae and Poaceae, showed 1,500-year moisture cycles (Turney et al., 2004).

Pollen obtained from a marine sediment core located 60 km west of the Cape Range Peninsula in semi-arid north-west Australia provides the only information about the late Quaternary vegetation in this region (van der Kaars and De Deckker, 2002). Mean annual rainfall today is between 200 and 300 mm. The site lies near the southern limit of the Australian summer monsoon today. Williams et al. (2009b) refined the interpretation of the pollen record from this core using the transfer functions of van der Kaars et al. (2006). The lowest mean annual precipitation and highest mean maximum temperatures in the period before the LGM occurred from 35 to 33 ka. The interval between 33 and 20 ka was the driest period of the 100 ka record, with almost no summer rain, in accord with previous suggestions that the summer monsoon failed to reach north-west Australia at this time (De Deckker et al., 2002; van der Kaars and De Deckker, 2002). Coldest temperatures occurred between 26 and 24 ka (van der Kaars et al., 2006). Conditions became slightly wetter and warmer from 20.4 to 14.2 ka and distinctly wetter after 14.2 ka, a change also seen at Lynch's Crater in north-east Queensland. In addition, the reconstructed late Pleistocene discharge record of the Fitzroy River and the onset of high lake levels in Lake Gregory in north-west Australia led Wyrwoll and Miller (2001) to conclude that the summer monsoon had become active there by 14 ka. This event reflects a global change in climate, with the abrupt return of the summer monsoon in the headwaters of the Blue and White Nile rivers at 14.5–14.0 ka (Williams et al., 2006c; Williams, 2012a). Furthermore, deMenocal et al. (2000) identified a sharp decline in eolian dust input from the Sahara into the Atlantic at this time, marking the start of what they termed the 'African Humid Period', although as Gasse et al. (2008) correctly observed, not all of Africa became humid at this time.

Playa lakes in the lower Darling Valley of semi-arid western New South Wales contain pollen showing that the plant communities growing in this region after the LGM (around 24–18 ka) were very different from those growing in the long interval (around 70–24 ka) before then (Cupper, 2005). Pollen preserved in eight middens of the stick-nest rat *Leporillus* in the arid northern Flinders Ranges of South Australia show that woodlands were widespread and the climate was wetter from 8.8 to 5.3 ka (McCarthy et al., 1996). During the preceding Pleistocene-Holocene transition, salt-tolerant plants were dominant. This may reflect continued aridity after the LGM or a change in rainfall seasonality at that time.

High-resolution U/Th ages from a speleothem in the Flinders Ranges, combined with carbon and oxygen isotopic analysis, showed high effective precipitation at around 11.5 and around 8–5 ka, with peak humidity at 7–6 ka (Quigley et al., 2010b), consistent with the stick-nest rat records of McCarthy et al. (1996). The present

climatic regime began about 5 ka ago and was characterised by more frequent ENSO events than the first half of the Holocene.

22.6.2 *Dunes and eolian sand plains*

Geomorphic evidence of arid zone landforms dated using cosmogenic isotopes shows that stony desert surfaces, or ‘gibber plains’, were present in Australia between 4 and 2 Ma, while the earliest recognisable sand dunes appear around 1 Ma ago (Fujioka et al., 2005; Fujioka et al., 2009). In contrast to the Sahara, which is only one-fifth sand, sand dunes and associated sand plains cover two-fifths of the present land area of Australia (Chapter 8; Figure 8.14). The younger dunes were episodically active during the late Pleistocene, including as far south as north-east Tasmania (Bowden, 1983; Duller and Augustinus, 1997), most notably at 73–66, 32–25, 22–18 (LGM) and 14–10 ka (Rhodes et al., 2004; Fitzsimmons et al., 2007a; Fitzsimmons et al., 2007b). Lomax et al. (2011) obtained a 380 ka record of dune activity in the western Murray Basin (Figure 22.2) based on ninety-eight OSL ages. Two major phases of dune accretion were at 72–63 and 38–18 ka, with minor phases at 14.5–13.5, 12–11 and 8–5 ka. They equated dune deposition with drier conditions and gaps in the record with wetter conditions. Hollands et al. (2006) dated linear dunes in the north-west Simpson Desert and concluded that there had been an approximately 160 km, or 1.5°, southward displacement of the sand-transporting wind system since the LGM. Along the eastern margin of present-day Lake Frome, transverse dunes were active at 111–106, 66–57 and 22–11 ka, and linear dunes were active by at least 66 ka and again at 43–28 ka, after which soils developed under a wetter climate. Clay pellets within several horizons of both transverse and linear dunes indicated salt influx and sporadic deflation of the lake floor, suggesting that local hydrologic conditions, rather than aridity, controlled dune activity (Fitzsimmons et al., 2007b).

The two periods with the greatest number of OSL ages are at around 20 ka, when sea level was 120 m lower than it is today and the land area of Australia was about 25 per cent greater (Figure 22.6), and at 14–10 ka, when temperatures were rising (Fitzsimmons et al., 2007a). The lack of ages between 20 and 14 ka seems to show that few deposits were preserved from that period, either because of subsequent reworking or because of minimal activity. We noted in Chapter 8 that dune sediments survive best if dune accretion is followed by a humid phase (Swezey, 2001; Swezey, 2003). We discuss source-bordering dunes in Section 22.6.5.

22.6.3 *Eolian dust*

The dust record preserved in marine cores collected off the east coast of Australia shows a threefold increase in dust flux during the LGM relative to the Holocene in temperate and tropical Australia (Hesse, 1994). The major present-day sources of dust

are the Lake Eyre Basin/Simpson Desert and the Murray-Darling Basin (Hesse and McTainsh, 2003), a pattern likely to have obtained in the past, with some modification. There was enhanced dust flux to the east and south of the continent between 33 and 16 ka (Hesse, 1994). Between 22 and 18 ka, the northern limit of the dust plume was 350 km, or 3°, north of its present limit (Hesse, 1994; Hesse et al., 2004), consistent with a northward shift of the high-pressure subtropical ridge (STR) during the LGM from its present summer location near 35° S. The STR separates the tropical easterly circulation from the mid-latitude westerlies (Hesse, 1994). Using present-day measurements of dust flux, McTainsh and Lynch (1996) estimated that dust activity in north-east Australia increased by 57 per cent and in the south-east by 52 per cent during the LGM. They pointed out that these were probably underestimates, given that they did not take into account increased sediment supplies to dust source areas at that time. McTainsh and Strong (2007) found up to 65 per cent organic matter by mass in modern dust, much of which was probably derived from biological crusts that normally protect soils and dunes from deflation. During the cold, dry LGM, it is probable that any existing crusts would also have been deflated.

Two exceptionally well-dated desert sites that span the Last Glacial Maximum are the Willandra Lakes in arid western New South Wales, particularly Lake Mungo, and the arid Flinders Ranges of South Australia, discussed in the next section. The chronology for late Pleistocene Lake Mungo is based on more than 200 radiocarbon and luminescence ages (Bowler, 1998; Bowler and Price, 1998) and shows that wind-blown dust began to accumulate in the lunettes on the eastern side of Lake Mungo and associated lakes from around 35 until around 16 ka, with a peak during the LGM. Clay dunes and gypseous lunettes were developing on the downwind margins of seasonally fluctuating lakes in many parts of semi-arid south-east and south-west Australia during the LGM (Bowler, 1973; Bowler, 1976; Bowler, 1978b; Bowler and Wasson, 1984), when the major deflation of dry lake-beds in the arid zone was underway (Magee and Miller, 1998).

In two marine cores off the coast of South Australia that span the last 170 ka, Gingele and De Deckker (2005) documented three phases of enhanced dust flux into the ocean at around 74–70, around 45 and around 20 ka. Each of these episodes coincides with times of lake desiccation, dune accretion and sparse vegetation cover in south-central Australia and with times of minimum insolation in these latitudes (Croke et al., 1996; Gingele and De Deckker, 2005; Williams et al., 2009b).

22.6.4 Late Quaternary wetlands

Reworked wind-blown dust also comprises a significant component of the late Pleistocene fine-grained valley fills that have accumulated episodically since the last interglacial in the arid Flinders Ranges of South Australia (Williams et al., 2001; Williams and Nitschke, 2005; Williams et al., 2006a; Williams and Adamson, 2008;

Glasby et al., 2007; Haberlah et al., 2010a; Haberlah et al., 2010b). The most recent valley fill is around 35 ka near the base and 17–16 ka near the top of the sequence. The overall chronology is based on some 30 OSL ages and 100 AMS radiocarbon ages on shell, charcoal and plant debris. The alluvial clays contain unbroken shells of freshwater snails and ostracods, and were laid down in perennial wetlands (Williams et al., 2001) and, more locally, as slack-water deposits after major floods (Haberlah et al., 2010a). The alluvial clays themselves consist at least in part of reworked wind-blown dust or loess blown in from the west and deposited on slopes of the north-south aligned ridges of quartzite, limestone and shale that form the Flinders Ranges. The fact that this wetland persisted throughout the coldest, driest climatic interval in late Pleistocene Australia indicates a complex and counter-intuitive response to climatic change. Williams and Adamson (2008) have put forward a simple bio-geophysical model to explain the enigma of a late Pleistocene wetland in the arid Flinders Ranges during a time of peak regional aridity.

In the semi-arid Snowy Mountains of south-east Australia, the three most recent glacial advances have been dated using the cosmogenic nuclide ^{10}Be to 32 ± 2.5 , 19.1 ± 1.6 and 16.8 ± 1.4 ka, with periglacial activity concentrated between 23 and 16 ka (Barrows et al., 2001). The orographic snow-line was about 600–700 m lower, and the lower limit of periglacial solifluction was at least 975 m lower (Galloway, 1965b). Because this latter limit is roughly equivalent to the 10°C isotherm for the warmest month, the temperature in the warmest month was at least 9°C cooler than today (Galloway, 1965b).

Low summer temperatures would have reduced water losses from evaporation and transpiration in the Flinders Ranges and elsewhere. Miller et al. (1997) estimated that millennial-scale mean temperatures at low elevations in arid inland Australia between around 45 and 16 ka were at least 9°C lower than they were after 16 ka. They based their estimates on the temperature-dependent amino acid racemisation reaction (see Chapter 6) in radiocarbon-dated emu eggshells from the continental interior to reconstruct subtropical temperatures at low elevations over the last 45 ka. They inferred a sharp change at around 16 ka, followed by rapid warming. These temperature data suggest that lower temperatures and reduced evaporation may have been factors in the persistence of wetland ecosystems in the Flinders Ranges (Williams et al., 2001; Williams et al., 2006a).

Low glacial concentrations of atmospheric carbon dioxide would have favoured grasses at the expense of trees. Trees act as natural groundwater pumps, so fewer trees, especially the River Red Gum, *Eucalyptus camaldulensis*, would lead to rising groundwater levels. The increase in dust flux was a function of greater aridity, reduced plant cover, increased wind velocity and wind gustiness (Maher et al., 2010; McGee et al., 2010). The grass cover on the hillsides provided an efficient dust trap for loess blown in from dunes, dry playa lakes and from the exposed continental shelf during times of glacially lowered sea level. A loess mantle on the slopes would increase

infiltration and reduce run-off, leading to an increase in base flow along the valley floors. Weaker summer monsoon precipitation would mean fewer extreme erosive events, allowing fine-grained dust mantles to be gradually removed from the slopes and to accumulate along the valley floors. A lower winter cloud base and gentle winter rains would facilitate progressive deposition of the reworked eolian clays and development of the late Pleistocene valley fills. The LGM was not uniformly arid, and at one locality in the central Flinders Ranges, up to twenty individual flood events are preserved as slack-water deposits (Haberlah et al., 2010a; Haberlah et al., 2010b).

There was a rapid change in environmental conditions at around 17–16 ka, with cessation of valley-fill formation and widespread vertical incision down to bedrock, followed by gully erosion that continues to this day. Singh and Luly (1991) analysed pollen from Lake Frome. They found that the treeless conditions of the LGM gave way to a flora dominated by *Eucalyptus* and *Callitris*, followed after 17 ka by a decline in *Callitris* and a more gradual decline of *Chenopodiaceae*. That these events were part of a global change in climate is suggested by the apparently synchronous retreat of mid-latitude mountain glaciers in both hemispheres from 17 ka onwards, which have been attributed to synchronous warming in those latitudes (Schaefer et al., 2006). Blue Lake in the Snowy Mountains was ice-free by 15.8 ka, indicating rapid deglaciation (Barrows et al., 2001).

22.6.5 Rivers and source-bordering dunes

The longest rivers flowing in Australia today are those like the Murray-Darling that rise in the Eastern Highlands and flow west or south-west across semi-arid plains, where much of their flow is lost in seepage and to evaporation during floods. In the south, snow-melt is an important contributor to these rivers, but in the north, they depend on summer rainfall. In the aptly named Riverine Plains in semi-arid south-east Australia, there is a series of large former channels that range from large and sinuous to large and straight (Chapter 10, Figure 10.9). In a classic study of the former channels in the lower Murrumbidgee Valley, Schumm (1968) concluded that the large, sinuous suspension-load channels had formed under a wetter climate with bank-full discharges several times those of today. The linear channels contained a coarse channel fill and fell into the category of bed-load channels. Schumm (1968) concluded that they were active under a more seasonal flow regime characterised by episodically very high discharge from more sparsely vegetated headwaters.

Bowler (1978a) expanded on this work and obtained a detailed radiocarbon chronology from charcoal within channel fill sediments. He found that during the LGM (21 ± 2 ka), the rivers flowing from the Eastern Highlands were transporting and depositing a coarse load of sand and gravel across the Riverine Plains until around 15 ka. With the advent of luminescence dating, it proved possible to obtain reliable ages for discrete phases of widespread fluvial aggradation at around 35–25 and

20–14 ka, during times of late Pleistocene flood discharges that were at least seasonally much higher than they are today (Page et al., 1991; Page and Nanson, 1996; Page et al., 1996; Ogden et al., 2001; Page et al., 2001; Bowler et al., 2006).

We saw in [Chapter 10](#) that there were close links between late Pleistocene snow-melt and river discharge and sediment deposition in south-east Australia (Williams et al., 2009b). Barrows et al. (2001) obtained ^{10}Be cosmogenic nuclide ages of 32 ± 2.5 , 19.1 ± 1.6 and 16.8 ± 1.4 ka for the three youngest glacial advances in the semi-arid Snowy Mountains of south-east Australia and ages between 23 and 16 ka for periglacial deposits in that region. A combination of sparse vegetation in late glacial times, together with extensive slope mantles formed by periglacial solifluction, would have meant that there was an abundant supply of coarse debris to rivers during the spring snow-melt. Further north beyond the limits of glacial and periglacial processes, the sparse plant cover would have contributed to high rates of run-off and a coarse sediment load, as evident in the large channels out on the alluvial plains (Williams, 1984e; Williams, 2000b; Williams, 2001a). An additional factor contributing to high LGM run-off was the influence of lower temperatures in the uplands, which could have at least doubled the run-off coefficient in the Snowy Mountains (Reinfelds et al., 2014). Once the climate became warmer and wetter during the very late Pleistocene and early Holocene, the plant cover became denser and soils began to form, leading to a change from bed-load to suspension-load channels. The decline in precipitation over the past 5 ka led to progressively smaller channels (Williams et al., 2009b).

There is also a direct association between Pleistocene river activity and source-bordering dune formation in central and south-east Australia ([Chapter 8](#)). Source-bordering dunes are common in Australian deserts wherever there is a regular supply of alluvial sand transported by seasonally flowing streams (Wasson, 1976; Bowler, 1978a; Bowler, 1978b; Williams et al., 1991a; Nanson et al., 1995; Page et al., 2001; Maroulis et al., 2007; Cohen et al., 2010a). There appear to be three prerequisites for the formation of fluvial source-bordering dunes. A regular supply of bed-load sands brought in by rivers that dry out seasonally, leaving their sandy point-bars and sandy channel beds exposed to deflation, is the first requirement. Strong, unidirectional winds are needed to move the channel sands and form a linear, or parabolic, dune. Sparse or absent riparian vegetation is needed for unimpeded sand movement out of the channel through deflation. The first condition is the most important, because without regular replenishment of the alluvial sand supply, the dunes will be unable to develop and propagate downwind.

Cohen et al. (2010a) have established a comprehensive chronology of fluvial and eolian sediments in the lower Cooper Creek region of central Australia based on fifty-seven TL and six OSL age estimates. Cooper Creek originates from the confluence of the Thompson and Barcoo rivers, both of which rise in the Eastern Highlands of Australia, where they are fed by tropical summer rainfall. The Cooper then flows

towards Lake Eyre in the heart of the arid zone, losing most of its discharge en route. Cohen et al. (2010a) concluded that enhanced flows were evident in the Cooper at 120–100 ka, 85–80 ka, 28–18 ka and during the early to mid-Holocene. The first three of these phases were associated with the more or less synchronous formation of source-bordering dunes. The dated stratigraphic profiles of the dunes indicate formation by vertical accretion of sediment, with little sign of downwind accretion and no evidence of long-distance transport of dune sand. Particularly interesting is the strong evidence of enhanced river flows immediately before and during the Last Glacial Maximum – a time widely considered as having been both very cold and very dry throughout Australia.

Maroulis et al. (2007) had earlier obtained a long luminescence chronology of alluvial sand transport along Cooper Creek from Marine Isotope Stages (MIS) 8 to 3, showing a trend of progressively declining discharge during that time. Peak source-bordering dune activity dated to late MIS 5 (around 85–80 ka) and mid-MIS 3 (50–40 ka), after which the flood-plains became mantled with mud and dunes became islands in a sea of mud, much as in the Gezira alluvial plain of central Sudan. Nanson et al. (2008) dated the alluvial deposits in the lower 500 km of Cooper Creek and found multiple episodes of enhanced flow during the last quarter of a million years. There was a progressive reduction in discharge until aridity became severe around 40–35 ka and source-bordering dunes ceased to form.

In northern Australia, the sediments associated with two relict plunge pools indicate extreme floods just before the LGM and during the early to mid-Holocene, with flood discharges up to five times greater than those of any floods in the last 40,000 years (Nott et al., 1996; Nott and Price, 1999). Extreme rainfall can therefore occur during both cold and warm climatic phases.

In the upper Shoalhaven River Valley in south-east Australia, Nott et al. (2002) dated four sets of paired alluvial terraces spanning about the last 500 ka. They found a lack of fluvial activity during the cold, dry LGM and an overall decrease in fluvial activity after 60 ka, consistent with an increase in regional aridity.

22.6.6 Lake fluctuations

The salt lakes of central and western Australia originated as a well-integrated Late Cretaceous to early Cenozoic drainage system that had become disrupted by the mid-Miocene as a result of successive tectonic movements (Bunting et al., 1974; Van de Graaff et al., 1977; Salama, 1997; Zheng et al., 1998; Zheng et al., 2002). The resulting lakes dried out progressively, some attaining peak aridity by around 500 ka (Zheng et al., 1998), when gypsum began to be widely deposited. Optical dating using green-light stimulated luminescence and electron spin resonance dating of relic gypsum dunes associated with two such lakes in south-west Australia showed intensified eolian deflation during the last glacial period, lunette construction during

the arid LGM and a return to high lake levels during the early to mid-Holocene (Zheng et al., 2002).

A 300 ka record from the Gregory Lakes in north-west Australia shows very high lake levels at around 300, around 200 and around 100 ka, or broadly coeval with MIS 9, 7 and 5 (Bowler et al., 2001). Dunes formed on the lake floors during drier phases dated to around 230 and around 70 ka. The largest lake (6,500 km²) was at around 300 ka, and later lakes were smaller, indicating increasing aridity over time, much as in the Sahara. These lakes were fed from the summer monsoon. An increase in the frequency of high-magnitude rainfall events would have helped generate these Quaternary mega-lakes. Pack et al. (2003) analysed the carbon isotopic composition of bulk soil organic matter from two sediment units in the Lake Gregory region spanning roughly the last 200 ka. They found a shift from mainly C₃ to mainly C₄ plant communities over the last 120 ka. For woodland to decrease and grassland to increase requires a decrease in rainfall and an increase in rainfall seasonality. They concluded that there had been a late Quaternary increase in aridity, consistent with the observations of Bowler et al. (2001).

Lake Eyre has the most detailed record of Upper Pleistocene fluctuations of any desert lake in Australia (Magee et al., 1995; Croke et al., 1996; Magee, 1998; Magee and Miller, 1998; DeVogel et al., 2004). The Lake Eyre Basin occupies one-sixth (1.3 million km²) of the present land area of Australia and thus contains a good portion of Australia's late Quaternary environmental record. It is fed mainly by the Cooper and Diamantina rivers, which flow from the Eastern Highlands and are fed primarily by the summer monsoon (Bonython and Mason, 1953; Kotwicki, 1986). At present, it is a vast saltpan and at its lowest point is 15 m below sea level. In 1974, it attained its highest historical level of -9.5 m. During La Niña years, when the seas east and north of Australia are warmer than average at the surface, precipitation is above average in northern and eastern Australia, and run-off into Lake Eyre leads to temporary flooding (Kotwicki and Isdale, 1991; Kotwicki and Allan, 1998).

The chronology that underpins the history of high and low lake stands over the past 130,000 years is based on three independent dating methods (radiocarbon, uranium-series disequilibrium and luminescence – both TL and OSL) supplemented by amino acid racemisation (AAR) dating of the egg-shells of emus and of their near relative, the now extinct giant ratite *Genyornis* found around the former lake margins (Magee and Miller, 1998). Lake Eyre attained its maximum level around 130–110 ka, with progressively lower levels thereafter, at 95–80 and 65–62 ka (Magee, 1998). Forty AMS radiocarbon dates show that the lake was dry between 35 and 10 ka (Magee and Miller, 1998), but these ages are in conflict with two TL ages obtained by Nanson et al. (1998) that suggest high beach levels during this time. Furthermore, the AAR ages obtained by Magee and Miller (1998) from these same beach ridges do not support the TL ages. Pending future work, this issue is best left open. During the last interglacial

(around 125 ka), Lake Eyre reached a level of +10 m, covered about 35,000 km² and, including the water in the Lake Frome complex, had a volume of about 430 km³, in contrast to the historic maximum of only 30 km³ (DeVogel et al., 2004).

The LGM in Australia was a time when the lakes in the seasonally wet tropics of northern Australia were mostly dry (English et al., 2001) and when Lake Eyre in central Australia was dry and its bed was being lowered by deflation (Magee et al., 1995; Croke et al., 1996; Magee and Miller, 1998; Magee et al., 2004). However, the interpretation of the lake level fluctuations is not as straightforward as it might seem. At any one time, the lake may have been fed in part from surface run-off, in part from groundwater influx and in part from direct precipitation, with the relative proportions of each varying over time. Another complicating factor concerns the role of groundwater-limited deflation in controlling lake basin shape, volume and subsequent shoreline levels during ensuing lake transgressions. If the long-term export of sediment through deflation exceeded the long-term replenishment of lake floor sediment, successive lake levels would be lower, although the net input of water might not have diminished. During the course of several glacial-interglacial cycles, with deflation especially active during the drier phases, the cumulative deepening of the lake floor may result in progressively lower lake shorelines, giving the possibly misleading impression that earlier interglacials were hydrologically more effective than later ones (Williams, 2001a). A way to test this is to use independent evidence to reconstruct past changes in precipitation. To this end, Johnson et al. (1999) analysed the carbon isotopes in fossil emu and *Genyornis* eggshells from around Lake Eyre. They found significant changes in the proportions of C₄ to C₃ grasses over the last 65 ka. They concluded from their analyses that the Australian monsoon was most effective between around 65 and around 45 ka, least effective during the LGM and moderately effective during the Holocene, all of which supports the reconstructed lake levels.

Bowler et al. (2011) have refined their earlier interpretation of the Willandra Lakes in semi-arid western New South Wales. These lakes function as a cascading system (Chapter 11, Figure 11.6) of lakes fed from Willandra Creek, a distributary channel of the Lachlan River, which has its headwaters in the Eastern Highlands (Chapter 11, Figure 11.3). Each lake basin reacted individually to fluctuations in discharge from the parent river, with some filling faster and drying out sooner than others. Some lakes were becoming saline at the same time that others were deep and fresh. Such a time-transgressive response to regional climatic changes is reminiscent of the late Pleistocene lakes fed by the Okavango River in Botswana.

Figure 22.9 is a synthesis of previous discussion and shows Australia and New Guinea during the LGM, when the sea level was about 120 m lower than today, temperatures were up to 8°C cooler and ice was present on the highest peaks. Figure 22.10 provides a vivid contrast, with warmer and wetter conditions prevalent across the region.

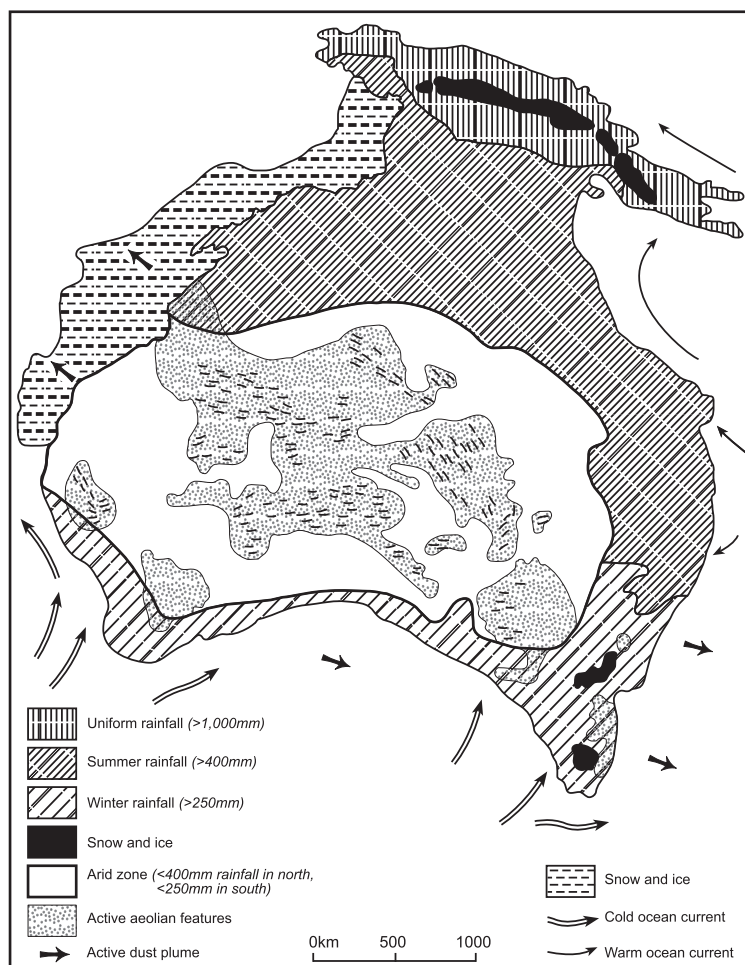


Figure 22.9. Australia and New Guinea during the LGM. (After Williams, 2001a, fig. 1.3A.)

22.7 Pleistocene extinctions of the Australian megafauna

The late Quaternary Lake Eyre Basin environmental record has also been at the heart of still unresolved debates over the role of prehistoric humans on faunal extinctions, on vegetation cover and on the strength of the summer monsoon. The disappearance of the megafauna some 50,000 years ago, including the giant flightless bird *Genyornis*, which, as noted in the previous section, was somewhat similar to the modern emu, coincides very broadly with what is presently known to be the first appearance of prehistoric humans in Australia (Roberts et al., 1994; Roberts et al., 2001; Bowler et al., 2003; Veth et al., 2009). Three main hypotheses have been proposed to account for the demise of the Australian megafauna: climatic change, ‘overkill’ as a result of hunting and modification of habitat by burning.

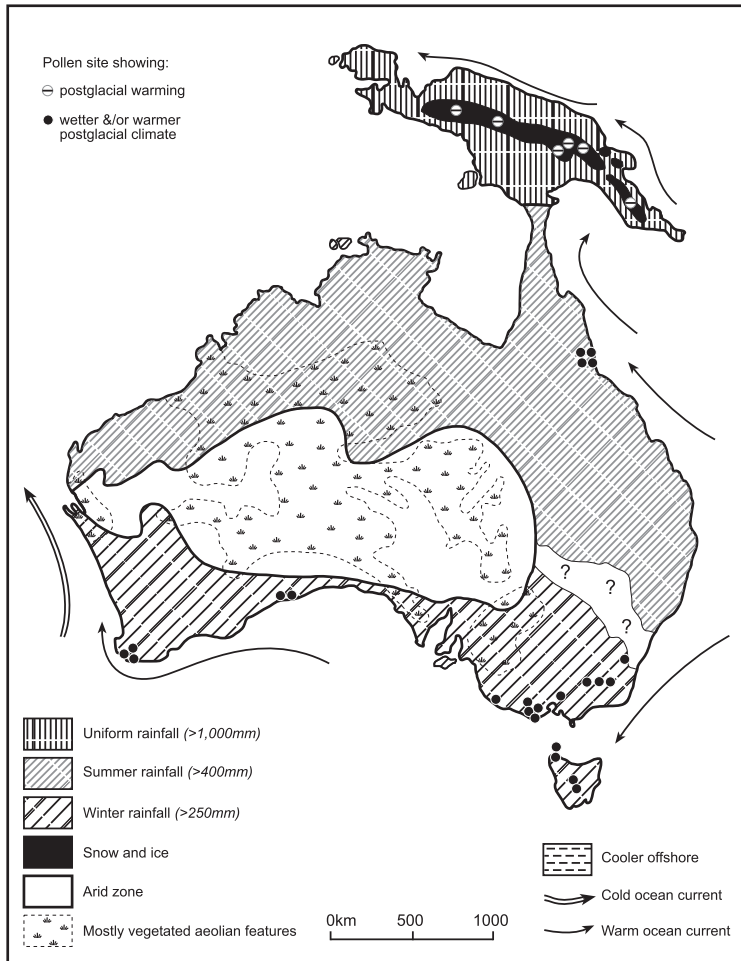


Figure 22.10. Australia and New Guinea during the early Holocene. (After Williams, 2001a, fig.1.3B.)

A number of Australian workers at present reject climatic change as a causal agent of extinction on the grounds that the animals had survived previous cold, dry glacial climates and that the extreme conditions of the Last Glacial Maximum some 20,000 years ago post-dated the extinctions (Roberts et al., 2001; Johnson, 2005). Decimation from hunting faces the immediate problem, discussed at length in Chapter 17, of the total lack of any discernible butchery sites. By default, the third hypothesis is accepted as the cause, with the rider that burning may have altered the local hydrology sufficiently to weaken the late Quaternary summer monsoon (Magee et al., 2004; Miller et al., 2005), a view opposed by certain climate modellers and palynologists (Lynch et al., 2007).

A serious weakness of the burning hypothesis stems from the sedimentary charcoal records obtained from 224 sites in and around Australia extending back to 70 ka, which show peak burning during wetter phases in both Australia and Indonesia and no increase in burning following the first arrival of humans on the continent at around 50 ka (Mooney et al., 2011). These authors rejected earlier arguments that humanly induced changes in fire regime and vegetation cover could have triggered regional climate change (Miller et al., 2005). On the contrary, they argue for a strong climatic control over biomass burning, with maximum burning during warm, wet intervals with the greatest plant cover and minimum burning during cold, dry intervals with least plant cover. In their statistical analysis of peaks and troughs in charcoal abundance, they discerned widespread millennial-scale fluctuations which have more in common with the Dansgaard-Oeschger climatic events recorded in Greenland ice cores (see Chapter 3) than they do with the record of climatic fluctuations preserved in the EPICA ice core record from Antarctica. However, they noted that the radiocarbon dating of many charcoal-bearing sites is not of adequate precision to demonstrate such a correlation with any real confidence. They also found significant geographical variation in the pattern of biomass burning in southern Australia during the Holocene.

A fourth working hypothesis may need to be considered and thoroughly tested (Williams et al., 1998; Wroe and Field, 2006). Because the extinct species of megafauna (such as *Procoptodon*, *Sthenurus* and *Diprotodon*) were primarily browsers rather than grazers, it is equally possible that a progressive and long-term increase in savanna grassland at the expense of forest and woodland may have contributed to a long-term decline in the numbers of browsing species, increasing their vulnerability to even minor impacts on their habitat, whether they were caused by humans or climate or both. There is growing evidence that the final demise of the megafauna occurred during a very dry interval. Van der Kaars et al. (2010) examined the pollen record from a deep-sea core off the coast of south-west Sumatra and found that the vegetation was most open during MIS 3, between around 52 and 43 ka, identifying this period as the driest of the last glacial, with the onset of cooler conditions starting at around 52 ka. Cohen et al. (2010b) showed that Lake Eyre was connected to Lake Frome for the last time at 50–47 ka, with a major shift to aridity thereafter, coincident with the demise of the megafauna and the arrival of humans.

It may be time to reconsider the impact of changes in climate. The evidence from desert lakes may help resolve whether the demise of the megafauna in Australia was a result of human impact (direct or indirect) or of climatic change. We know that the arrival of humans in Australia about 45 ka ago seems to coincide with a wave of faunal extinctions, so it might seem that humans were the cause of these extinctions. However, until recently, our knowledge of the climate at this time was sketchy. This lacuna is now being remedied. Cohen et al. (2010b) have shown that aridity set in soon after 45 ka in central Australia. Before then, the climate across the continent was much wetter and lakes were greatly expanded. Lake Frome immediately east of the

Flinders Ranges and a series of more northerly lakes were full, and overflowed into a much larger Lake Eyre at intervals until a final major transgression at 50–47 ka. The lake levels became progressively lower thereafter, with minor higher levels in Lake Frome at around 30, 17, 13, 5 and 1 ka. Drawing on the evidence from speleothems in caves located in the northern summer rainfall and southern winter rainfall zones, Cohen et al. (2010b) deduced that southern sources of rainfall contributed to run-off into Lake Frome during the 50, 30 and 17 ka lake transgressions. They also inferred that there was a tropical contribution to Lake Eyre via the Cooper and Diamantina rivers at 50–47 ka. The 13 ka and younger high levels of Lake Frome seem to be a result of northerly inputs of run-off from tropical sources. More of this type of work is needed to clarify the possible role of climatic change in the extinction of the Australian megafauna.

22.8 Modelling changes in the Australian summer monsoon

There has long been a critical awareness of the influence of the Siberian high-pressure cell on the summer monsoon in northern Australia. However, less well-appreciated is the role that influxes of moist air from well south of the equator had on the strength of the summer monsoon in north-west Australia in particular. In an effort to resolve this issue, Wyrwoll et al. (2007, 2012) used regional atmospheric circulation models to explore the influence of changes in the earth's axial tilt and those of the orbital precession cycle on changes in wind direction and strength associated with synoptic conditions during the onset of the north-west Australian summer monsoon. The outcome of the model experiments showed that when the axial tilt is greater than it is today, there is a significant influx of air into the north-west of Australia from southern latitudes offshore, leading to enhanced summer monsoonal precipitation. The same is true of times of Southern Hemisphere perihelion.

22.9 Conclusion

Until about 120 million years ago Australia, Antarctica and Greater India formed a single supercontinent. The long-term desiccation of Australia was caused by its progressive separation from Greater India and from Antarctica. Greater India began to move away from Australia in the early Cretaceous around 120 Ma ago. The associated rifting and rift-margin uparching contributed to drainage disruption in the west of Australia. Initial rifting between Australia and Antarctica began around 90 Ma ago, and from about 45 Ma onwards, Australia moved north at a rate of 6–7 cm/year. As the eastern margin of the Australia plate moved north, it passed over one or more stationary hot spots in the mantle, resulting in volcanism and further uplift of the Eastern Highlands. Movement of Australia into tropical latitudes associated with dry subsiding air and semi-permanent high pressure led to accentuated aridity in central

Australia, which was first evident in the mid-Miocene some 15 Ma ago. Rainforest gave way to dry sclerophyll woodland, and a eucalypt flora adapted to frequent natural fires came into being and expanded across the continent. The trend to aridity occurred in stages, with the first stony tableland landscapes appearing at 4–2 Ma and the first desert sand dunes appearing by 1 million years ago. Large lakes shrank progressively, and from about 300 ka onwards, the inland lakes show a progressive reduction in size, associated with enhanced aridity, culminating in the cold, dry and probably windy Last Glacial Maximum some 20,000 years ago. Although the Australian flora is reasonably well-adapted to aridity, it is often highly sensitive to extreme cold. The two extremes of a glacial-interglacial cycle epitomise the full range of environmental changes that occurred on the continent during the Quaternary. During interglacials, the lakes were full and fresh, dunes were vegetated and stable, rivers were perennial, and rainforest and woodland expanded across the continent. By contrast, full glacial times were cold and generally dry, dunes were reactivated, lakes dried out or became saline, dust-storms were more frequent, grassland and shrubland replaced forests and woodlands, limited mountain glaciers and small ice caps occupied the highest uplands in the south-east and Tasmania, and previously perennial rivers became more seasonal or even ephemeral. The initial arrival of prehistoric humans around 45 ka ago coincides with a wave of faunal extinctions, as well as evidence of an increase in aridity. The jury is still out as to the precise causes of the faunal extinctions.

23

Historic floods and droughts

I did not know thee in the wilderness, in the land of great drought.
Hosea 13.5

The effort to control the health of land has not been very successful.
It is now generally understood that when soil loses fertility, or
washes away faster than it forms, and when water systems exhibit
abnormal floods and shortages, the land is sick.
Aldo Leopold (1887–1948)
A Sand County Almanac (1949)

23.1 Introduction

Anyone who has travelled widely and frequently in deserts may at some stage have experienced the full might and majesty of a desert flood. These rare events, once seen, are never forgotten. I retain vivid memories of the sudden and prolonged downpour on 5 August 1999 and of the ensuing torrential flow from the nearby mountains across the wide boulder-strewn valley floor we were crossing, which brought to an end the three-year drought in the Right Banner area of the Alashan Plateau of Inner Mongolia in arid northern China. The widespread floods in the northern uplands of Tunisia and Algeria, which culminated on New Year's Eve 1969, were associated with heavy rain as far south as the Saharan oases of Biskra and Toggourt – a boon to the local people despite the temporary discomfort. The floods of January 2007 in the arid Flinders Ranges of South Australia saw huge River Red Gums (*Eucalyptus camaldulensis*) transported westwards across the mountain piedmont to be dumped in great tangles on the surface of the flood-plain. Since the ephemeral stream channels were already incised more than ten metres into these plains, the flow velocity across the flood-plain must have been substantial. Considerable loss of life may ensue from these sudden floods. In early 1941, most of the local inhabitants of Kufra Oasis in the south-east

Libyan Desert had taken refuge in nearby low-lying caves to avoid the fighting, and perished from drowning during a sudden flash flood that submerged the caves. In similar vein, French legionaries bivouacked on the sandy bed of a normally dry wadi in the Algerian Sahara were drowned by a sudden flash flood that originated in the adjacent Atlas Mountains and of which they were entirely unaware (Sparks, 1972, p. 331). After that, the French Army changed their standing orders to forbid bivouacs in desert wadis.

In their detailed analysis of the interactions between desertification and climate, Williams and Balling (1996, p. 17) stated that drought 'denotes a regional deficiency in soil moisture which may be caused by a combination of lower than normal precipitation and higher than average evapotranspiration'. However, they also pointed out that drought has many wider connotations not covered by this definition. For example, in the UNESCO-WMO report on *Hydrological Aspects of Drought*, the rapporteurs noted that 'drought is generally viewed as a sustained and regionally extensive occurrence of below average natural water availability, either in the form of precipitation, river run-off or groundwater' (Beran and Rodier, 1985, p. 1). Depending on when it occurred, a rainfall deficit might have little impact on local pastoralists but a major impact on local cultivators. Likewise, a reduction in shallow groundwater recharge might have no immediate impact but an adverse one in future years. These differing aspects of drought are apparent in such concepts as 'agricultural drought', 'hydrological drought' and 'meteorological drought', and they are analysed in detail elsewhere (WMO, 1975; WMO, 1990; Beran and Rodier, 1985). To these we might add 'edaphic drought', which is when the soil has become so eroded and degraded that it cannot absorb much moisture, and the resulting extreme run-off leads to flash floods, as indicated in the quotation from Aldo Leopold (1949) at the start of this chapter. One consequence of widespread and prolonged drought is the reduction of global primary production. Zhao and Running (2010) estimated that the large-scale droughts during the warm decade 2000–2009 reduced global net primary production by 0.55×10^{15} g of carbon.

The aim of this chapter is to examine some of the more important factors that have controlled the magnitude and frequency of historic floods and droughts and to investigate the degree to which recent human activities may be contributing to floods and droughts in the desert world and its margins. Along the semi-arid margins of deserts, such extreme floods are more frequent and generally more widespread than in the hyper-arid areas; they also appear to be occurring more often in certain areas than in the time since flood records were first recorded. For example, in the Mojave Desert of the south-west United States, in the lower Narmada Valley of north-central India, and in the Fitzroy and Katherine rivers of northern Australia, recent flood levels have reached heights above present river level last attained several thousand years earlier (Baker et al., 1985; Ely et al., 1993; Baker et al., 1995; Ely et al., 1996).

In [Chapter 1](#), we defined deserts and desert margins as including the hyper-arid, arid, semi-arid and dry subhumid regions of the world, which together make up about 47 per cent of the total land area of the globe ([Chapter 1](#), [Table 1.1](#)). The dry subhumid areas include the drier parts of the seasonally wet tropics (see [Chapter 1](#), [Figure 1.1](#)) and fall within the domain of the summer monsoon ([Figure 23.1](#)). Year to year variations in the summer monsoon reflect fluctuations in the heat budget over this region modulated through the Hadley Circulation, which was discussed in [Chapter 2](#). These fluctuations are in part controlled by the relative temperature differences between land and sea in early summer. In the case of Asia, these fluctuations are also linked to variations in the strength of the winter anticyclone over Siberia, which controls the strength of the winter monsoon and the extent of snow cover over central Asia (Diaz and Markgraf, 1992; Wang, 2006). Another major cause of interannual variation stems from changes in sea surface temperatures over the equatorial Pacific Ocean and adjacent Indian Ocean that are associated with the Southern Oscillation, to which we now turn.

23.2 El Niño-Southern Oscillation (ENSO) events and historic floods and droughts

Sir Gilbert Walker first recognised and defined the Southern Oscillation when he was seeking more effective ways of predicting Indian summer monsoon rainfall (Walker, 1924). The Southern Oscillation is a measure of the surface atmospheric pressure differences between the western and eastern limbs of the equatorial Pacific. Walker observed that when the pressure off the coast of Peru is below average, that at Jakarta is above average, and vice versa ([Figure 23.1](#)). The difference is expressed as an index termed the Southern Oscillation Index (SOI), which is now generally taken to be the pressure difference between Tahiti in the central equatorial Pacific and Darwin in tropical northern Australia (Glantz et al., 1991; Diaz and Markgraf, 1992; Allan et al., 1996; Williams and Balling, 1996; Power et al., 1998; Grove and Chappell, 2000; Diaz and Markgraf, 2000; Peel et al., 2002). [Figure 23.1](#) shows the main areas influenced by the Southern Oscillation, which coincide very broadly with the domain of the summer monsoon. The changes in atmospheric circulation associated with the Southern Oscillation are termed the Walker Circulation.

The relationships calculated by Walker have stood the test of time (Diaz and Markgraf, 1992), although they were at first criticised. When the SOI is strongly negative, droughts tend to be common in certain parts of the world and are often synchronous in regions as far apart as the Ethiopian Highlands, peninsular India, eastern China, northern Thailand, Java, north-east Brazil and eastern Australia. Conversely, during years when the SOI is strongly positive, major floods are common and are usually synchronous in these same regions. Peruvian fisherman have long recognised that around

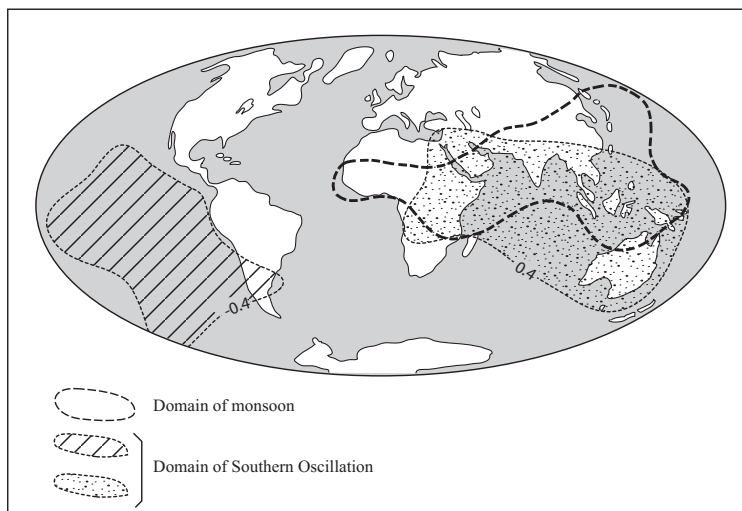


Figure 23.1. Region influenced by the summer monsoon and the two key regions of the Southern Oscillation. During El Niño, or ENSO, events (negative Southern Oscillation Index), surface atmospheric pressure is above normal in the stippled area and below normal in the hatched area. The opposite pattern prevails during La Niña, or anti-ENSO, events (positive SOI). For example, when surface atmospheric pressure is below-average off the coast of Peru, it is above-average in the area encompassing India, southern China, East Africa, Indonesia and Australia, and drought is common in these localities. The mapped areas (showing statistically significant correlations greater than $+0.4$ or -0.4 , at the 95% confidence level) reflect the difference between annual surface atmospheric pressure at Jakarta, Indonesia, and global mean sea level pressure. (Modified from Whetton et al., 1990, and Williams et al., 2006b.)

December in certain years, the normally cold and nutrient-rich waters offshore are replaced by a tongue of warmer water, and the anchovy fisheries fail. They term these years ‘El Niño years’, and we now have a reasonably accurate record of such events spanning the last 500 years (Whetton et al., 1990; Diaz and Markgraf, 1992; Whetton et al., 1992; Whetton and Rutherford, 1994; Allan et al., 1996; Whetton et al., 1996; Diaz and Markgraf, 2000; Ortlieb, 2004). (El Niño is Spanish for ‘little boy’ and refers to the Christ child, because December is the usual month in which an El Niño event starts). Such years are accompanied by severe floods in normally arid coastal Peru and are years of negative SOI, with lower-than-average surface atmospheric pressure offshore. Years when the waters off Peru are colder than average are termed ‘anti-El Niño years’ or, more simply, ‘La Niña’ years. The phrase El Niño-Southern Oscillation (or ENSO) event is a concise way of indicating a year marked by an El Niño event and a negative SOI. The two phenomena are closely related but are not synonymous, given that one refers to surface atmospheric pressure (Southern Oscillation) and the other to sea surface temperature anomalies (El Niño events).

Table 23.1. *Data sets used in the compilation of time series shown in Figure 23.2. (The means and standard deviations were calculated for the 1911–1960 reference period; adapted from Whetton et al., 1990, table 1.)*

Data set	Record length	Mean	Standard deviation	Units
Southern Oscillation Index	1870–1986	0.18	0.79	
North China Rainfall Index	1970–1979	3.07	0.44	
India rainfall	1871–1985	992.2	82.6	mm
Krishna River discharge	1901–1960	57.1	14.5	10 ⁶ MI

Comparison of the 500-year record of wet and dry years in China (Anon., 1981) and the corresponding flood record of the Nile with the updated El Niño record from Peru revealed some previously unsuspected correlations. Whetton et al. (1990) carried out time series analysis of rainfall and river flow in India, wet and dry years in China, and corresponding values of the SOI (Tables 23.1 and 23.2; Figure 23.2). These authors were the first to show that in eastern China during the period 1870–1979, years of strongly negative SOI coincided with a poor monsoon in the three key agricultural regions shown in Figure 23.3.

In much of peninsular India since at least the year 1871, years of severe drought are generally (but not invariably) years of strongly negative SOI (Figure 23.2). Table 23.1 specifies the data sets used by Whetton et al. (1990) in compiling the time series shown in Figure 23.2 and Table 23.2. (An update of this analysis would be

Table 23.2. *Statistically significant correlations between China rainfall, India droughts, Java tree rings, Nile flood height and El Niño occurrences in Peru for different time intervals between 1740 and 1984. (Nile floods and Java tree rings correlated significantly during the 1750s and from 1870 to 1929 [0.01]; adapted from Whetton et al., 1992, fig. 25, with supplementary data from Whetton and Rutherford, 1994.)*

Significant correlation	Locality A	Locality B	Time interval	Significance
1	El Niño	North China rainfall	1770–1879	0.08
2	El Niño	Nile floods	1770–1869	0.08
			1880–1984	0.001
3	El Niño	India droughts	1770–1869	0.03
4	El Niño	Java tree rings	1770–1984	0.01
5	Nile floods	North China rainfall	1870–1984	0.01
6	Nile floods	India droughts	1770–1879	0.05
			1880–1984	0.01
7	India droughts	North China rainfall	1870–1984	0.001

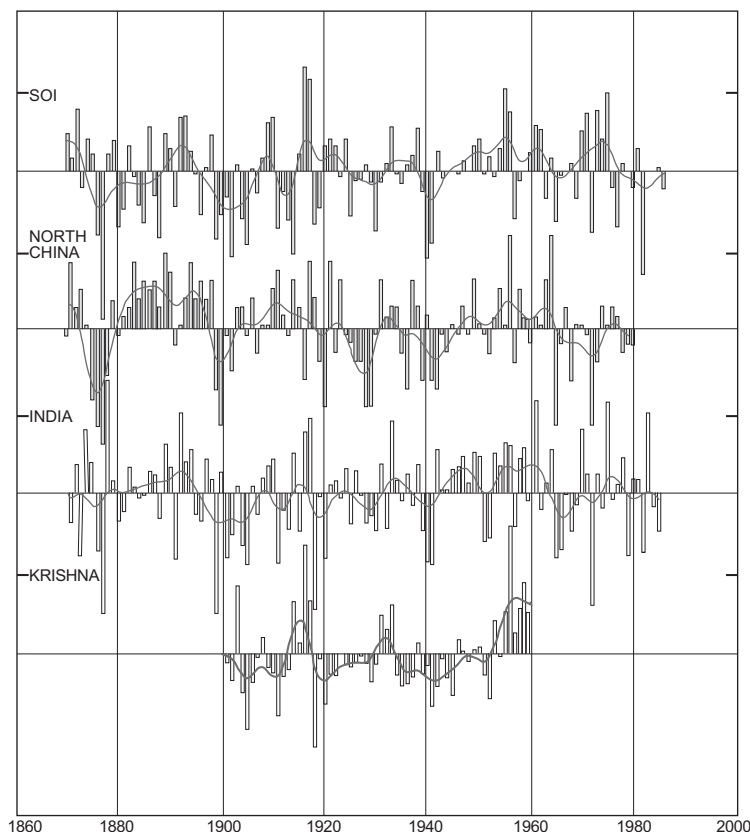


Figure 23.2. Time series representation of the Southern Oscillation Index (1870–1986), the annual rainfall index for China (1870–1979), rainfall over India in mm (1871–1985) and discharge in the Krishna River, India, in millions of Ml (1901–1960). Each series has been standardised with respect to the means and standard deviations given in Table 23.1 for the 1911–1960 reference period. Each interval on the vertical axis represents two standard deviations. The solid curved line shows a smoothed version of the time series, constructed using a ten-year Gaussian filter, which suppresses frequencies of less than ten years. (After Whetton et al., 1990, and Williams and Balling, 1996, fig. 5.4.)

informative, especially given the unusually warm global land and sea temperatures of the last twenty or so years, but to the best of the author's knowledge one has not been carried out). There is also good evidence that since the 1970s, the links between Indian monsoonal rainfall and ENSO events have weakened (Kumar et al., 2006; IPCC, 2007a; IPCC, 2007b), perhaps reflecting changes in the Indian Ocean Dipole (Ashok et al., 2001) or perhaps reflecting the influence of warmer sea surface temperatures in the northern Indian Ocean and the Arabian Sea. We discuss the Indian Ocean Dipole in the next section.

Godley (2002) extended this type of analysis to the drier parts of Thailand, where he found that drier-than-average years in northern and central Thailand were El

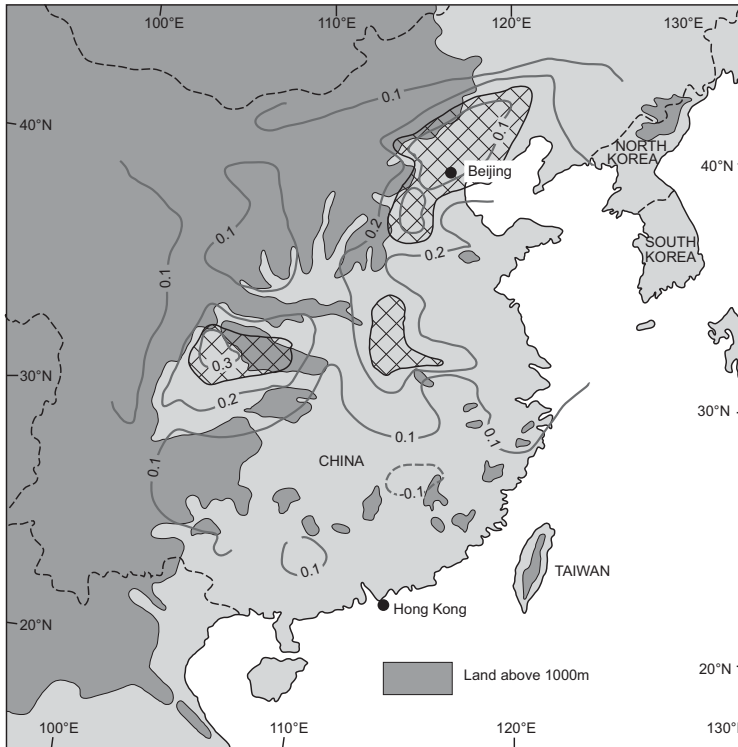


Figure 23.3. Map showing the correlation between the annual rainfall index for China and the Southern Oscillation Index for June, July and August for the 1870–1979 period. Correlation coefficients greater than 0.25 (cross-hatched areas) are significant at the 1% level, after allowing for the reduction in degrees of freedom due to persistence, although this effect is quite small. (Modified from Whetton et al., 1990, fig. 3.)

Niño years, and floods occurred during La Niña years (Figure 23.4). In Java, drier-than-average years result in slower-than-average growth among teak trees (*Tectona grandis*), so that the long tree ring record for the 1852–1929 period collated by Berlage (1931), published by de Boer (1951) and reanalysed by Murphy and Whetton (1989) can be compared with the SOI. Figure 23.5 shows that years characterised by narrow tree rings in Java were also years of negative SOI. Tapper (2002) has shown that during the 1877–1998 period, ENSO events coincided with both droughts and fires in Indonesia (Table 23.3). There is also a close correlation between years of narrow tree rings in Java and years when the Nile gauge at Roda in northern Egypt showed significantly below-average flows in the Nile at Cairo (Figure 23.6), indicating failed summer rains in the Ethiopian headwaters of the Blue Nile. The intervals 1737–1800 and 1825–1903 were chosen because they reflect little or no interference from dams and other river regulation structures with flood gauge readings.

Figure 23.7 summarises much of the preceding discussion into a single diagram that shows statistically significant correlations, or ‘teleconnections’, between rainfall

Table 23.3. Documented occurrences of El Niño-Southern oscillation events in relation to droughts and fires in Indonesia (1877–1998). (Derived from Tapper, 2002, table 2.1.)

Year(s) and ENSO	Drought severity and broad location	Fires and broad location
1877 Strong ENSO	Java (no data elsewhere)	Central Kalimantan
1888 Strong ENSO	South – severe	
1896 Strong ENSO	South – not severe	
1902 Weak ENSO	South and east – severe	
1911 Weak ENSO	Localised – not severe	
1914 Strong ENSO	East – severe	South and East Kalimantan
1918 Weak ENSO	South and east – moderate	
1930 Weak ENSO	East – moderate	
1940–1941 Strong ENSO	East – severe	
1940s–1950s	Lack of records	
1961 No ENSO	South and east – severe	
1963 Weak ENSO	South and east – severe	
1965 Strong ENSO	South and east – severe	
1967 No ENSO	South and east – moderate	
1969 Weak ENSO	South – moderate	
1972 Strong ENSO	South and east – severe	
1976–1977 Strong ENSO	South and east – severe	
1982–1983 Strong ENSO	Widespread – severe	East Kalimantan
1986–1987 Strong ENSO	South and east – moderate	All Kalimantan
1991–1994 Strong ENSO	Widespread – severe	Sumatra, Kalimantan, Java
1997–1998 Strong ENSO	Widespread – severe	Sumatra, Borneo generally

Strong ENSO = dry season SOI below –10

Weak ENSO = dry season SOI between 0 and –10

No ENSO = dry season SOI above 0

in India and China, river flow in India and the Nile, Java tree rings and ENSO events. Table 23.2 summarises these findings and shows the statistically significant correlations between China rainfall, Indian droughts, tree rings in Java, Nile flood height and El Niño occurrences in Peru for different time intervals between 1740 and 1984. Years of extreme drought were synchronous in north-east China, India, Ethiopia and south-east Australia in the years 1877, 1899, 1902, 1941, 1965 to 1966, 1972 and 1982 to 1983, and coincided with years of strongly negative SOI (Whetton et al., 1990). During the 1877 drought, some 10 million people died of starvation in China, as did more than 5 million in India and reputedly nearly one in two people in Ethiopia. Major floods were synchronous in these same regions during the years 1887, 1889 to 1890, 1894, 1916 to 1918, 1955 to 1956, 1964 and 1975, when the SOI was strongly positive (Whetton et al., 1990).

In a number of areas bordering deserts, including southern and eastern Africa, north-east Brazil, New Mexico, eastern and northern Australia, central India and

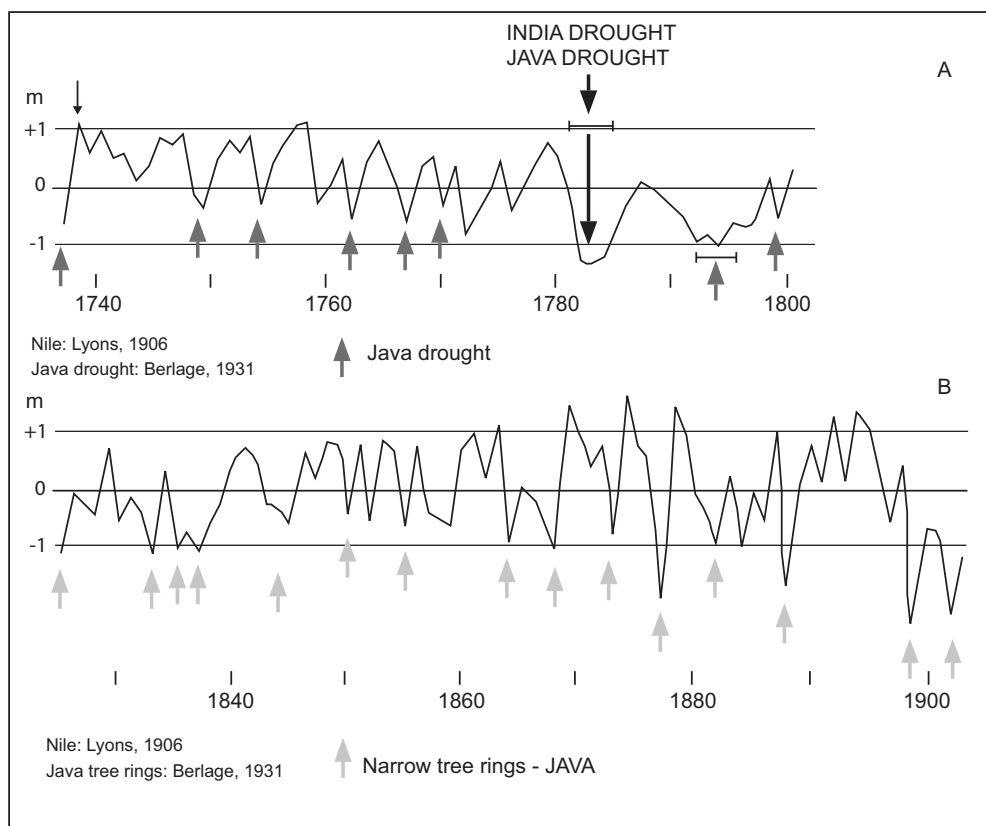


Figure 23.6. Nile River flood height at the Roda Gauge (1737–1903) showing correlation between droughts and/or years of narrow *Tectona grandis* (teak) tree rings in Java and years of below-average flow in the Nile. (Unpublished compilation by the late Dr Donald Adamson sent to the author in 1985, based on data obtained from Berlage, 1931, and Lyons, 1906, pl. XLIII, verified by the present author.) Variation from the mean values of maximum annual flood height readings at Roda is averaged for (a) the 1737–1800 period and (b) the 1825–1903 period.

north-east China, as we saw above, the incidence of wet and dry years is strongly influenced by the incidence of ENSO events (Kane, 1997; Peel et al., 2002). Many of the rivers in these areas are sensitive to small changes in rainfall, and ENSO events will amplify their already highly variable flow regimes (Adamson et al., 1987b; Kuhnel et al., 1990; Simpson et al., 1993a; Simpson et al., 1993b). The sensitive hydrological response of desert rivers to global sea surface temperature anomalies is an integral part of global environmental change, and it is likely to remain so in the future.

El Niño-Southern Oscillation (ENSO) events are a major source of rainfall variability both today and in centuries past, accounting for up to 50 per cent of rainfall variance in regions as widely dispersed as north-east Brazil, India, eastern China, eastern Australia, Indonesia, Thailand, the Nile Basin and southern Africa. The regions of



Figure 23.7. Statistically significant correlations (significance shown in brackets below) between China rainfall, Indian droughts, Java tree rings, Nile flood height and El Niño occurrences in Peru for different time intervals between 1740 and 1984. (1) is El Niño and northern China rainfall 1770–1879 (0.08); (2) is El Niño and Nile floods 1770–1879 (0.08) and 1880–1984 (0.001); (3) is El Niño and India droughts 1770–1869 (0.03); (4) is El Niño and Java tree rings 1770–1984 (0.01); (5) is Nile floods and northern China rainfall 1870–1984 (0.01); (6) is Nile floods and India drought 1770–1887 (0.05) and 1880–1984 (0.01); (7) is India droughts and northern China rainfall 1870–1984 (0.001). Nile floods and Java tree rings correlated significantly during the 1750s and from 1870 to 1929 (0.01). (Adapted from Whetton et al., 1992, fig. 25, with supplementary data from Whetton and Rutherford, 1994.)

annual pressure anomalies associated with the Southern Oscillation extend well beyond the two tropics in both hemispheres. However, the correlations between different localities, although statistically significant, do not account for all of the year-to-year variation in the strength of the summer monsoon, but they do account for a modestly important part of the interannual rainfall variability. Finally, at periodic intervals, for reasons we still do not understand, the spatial pattern of variation changes quite abruptly, so that two localities that were previously in phase suddenly cease to be so. A corollary to this is that future changes in the links between floods, droughts and ENSO events are to be expected and may offer surprises. Both of these conclusions accord with the views expressed throughout the three latest IPCC reports (IPCC, 2007a; IPCC, 2007b; IPCC, 2007c).

An important study by Power and Smith (2007), published soon after the three 2007 IPCC reports (IPCC, 2007a; IPCC, 2007b; IPCC, 2007c), has revealed that from 1977 to 2006, average values of the June–December SOI were the lowest on record, indicating a weakening of the Walker Circulation. At the same time, mean sea level atmospheric pressure at Darwin was the highest recorded, equatorial surface wind-stresses were at their weakest and tropical sea surface temperatures were the highest on record. They concluded that global warming must be considered in defining new ENSO indices and in assessing and using statistical correlations between ENSO events and climatic variations across the globe.

At present, the past offers us some valuable guidance as to which regions of the world are most influenced by ENSO, and this is unlikely to change very significantly in the future (Collins et al., 2010), although Vecchi et al. (2006) considered that the tropical Pacific atmospheric circulation would weaken, as would ENSO events. Indeed, the IPCC authors were unable to discern any significant change in the pattern of such events in space and time since observational records became available, and concluded that they could not predict likely future changes in the magnitude and frequency of ENSO events (IPCC, 2007a).

23.3 Historic floods, droughts and the Indian Ocean Dipole

Although part of the year-to-year and decadal-scale variation in the Indian monsoon is indeed linked to ENSO events, other influences are also at play. Saji et al. (1999) examined forty years of observations and were the first to identify what they termed a dipole mode in the tropical Indian Ocean, which soon became known as the Indian Ocean Dipole (IOD). They discovered that during years when the sea surface temperatures (SSTs) off Sumatra were unusually low, they were correspondingly high in the western Indian Ocean (termed the positive phase by Li and Mu, 2001). One outcome of these SST anomalies was drought in Indonesia coinciding with exceptionally heavy rainfall in East Africa. More than a decade earlier, Flohn (1987) had described the catastrophic rains of 1961 in tropical eastern Africa and the sudden increase in White Nile floods. This was also a year of pronounced IOD activity. The converse also applies, with drought in East Africa coinciding with low SSTs in the western Indian Ocean and with high SSTs and very wet conditions in Indonesia (termed the negative phase by Li and Mu, 2001).

Saji et al. (1999) also pointed out that because the dipole mode was strongly dependent on the state of the circulation associated with the summer monsoon, there would be interaction between the monsoon and the IOD. In the same issue of *Nature* in which Saji et al. (1999) had identified the IOD, Webster et al. (1999) elaborated on coupled ocean-atmosphere dynamics in the Indian Ocean, concluding that the 1997–1998 IOD was independent of ENSO, and Anderson (1999) observed that the dipole is reflected not only in SST variation but also in subsurface ocean temperatures in the

Indian Ocean. Later work by Ashok et al. (2003) confirmed that the IOD is indeed independent of ENSO. Depending on which mode of the IOD is operating, it may either strengthen or nullify the impact of ENSO on the Indian monsoon and associated rainfall (Ashok et al., 2001). Li and Mu (2001) extended the analysis and concluded that the IOD had a significant impact on atmospheric circulation and climate in North America and the southern Indian Ocean region, including Australia and southern Africa.

23.4 Longer-term records of ENSO activity

Other sources of proxy data provide a longer record of El Niño variability over the past 5,000 years and more. Moy et al. (2002) analysed a long lake record from Ecuador that was very sensitive to El Niño and reported an increase in El Niño activity from about 4,800 years ago onwards, after a long interval of some 5,000 years during which El Niño events were rare. They also observed that from about 1,600 years ago onwards, the frequency of El Niño activity has declined. Woodroffe and Gagan (2000) and Woodroffe et al. (2003) analysed the oxygen isotope ratios in *Porites* coral microatolls from Christmas (Kiritimati) Island in the central equatorial Pacific and obtained a high-resolution record of ENSO events for the last 3,800 years. They found that ENSO-related interannual variations in sea surface temperature and precipitation were less intense 3,800 to 2,800 years ago but became more pronounced 1,700 years ago, which is about the time that Moy et al. (2002) report that the onset of a decline in El Niño frequency in Ecuador occurred. Woodroffe et al. (2003) proposed that the amplification of ENSO events roughly 2,000 years ago could reflect stronger interactions between the Southern Oscillation and the Intertropical Convergence Zone, leading to stronger summer rainfall.

The magnitude and frequency of ENSO events during the Holocene show considerable variability. Cobb et al. (2013) obtained a 7,000-year record of ENSO events from fossil corals on the Christmas and Fanning islands in the tropical central Pacific. They found that ENSO activity during that time was highly variable, with no discernible trend, and that the relatively high twentieth-century variability was still within the range of past fluctuations. They concluded that it will prove hard to detect signs of anthropogenic, as opposed to natural, forcing of ENSO activity, given the large internal variability evident during the past 7,000 years. Bacon et al. (2010) completed a detailed study of late Holocene alluvial fan activity on the Muggins Mountains in the Sonoran Desert near Yuma in south-west Arizona and found evidence of widespread fan aggradation between 3.2 and 2.3 ka, which they attributed to rapid climate change and more intense ENSO events. They found no evidence of any historic reactivation of the alluvial fan surfaces, although the precipitation records did show numerous above-average rainfall events correlated with the SOI.

These findings from the geologically recent record of El Niño events reinforce the earlier conclusion that both the magnitude and frequency of ENSO events are

likely to change in the future in ways we cannot yet predict. The regions indicated in [Figure 23.1](#) as the ones most affected by ENSO during the last 100 and more years of meteorological records will remain susceptible in the future, with a very strong likelihood of increased variability in both temperature and precipitation (IPCC, 2007a). Nicholls (1989) has suggested that in regions strongly influenced by ENSO events, the plants and animals have developed adaptive strategies to cope with the rainfall variability.

The Nile provides an exceptionally long record of past floods and droughts. ENSO events are generally well-reflected in the last 500 years of Nile flow records (Ortlieb, 2004). In addition, low Nile flows were common between 950 and 1250 AD, roughly coeval with the Medieval Warm Period (MWP) in Europe. There were also severe droughts in East Africa during the MWP which caused major human migrations in this region but do not appear to have been in any way caused by anthropogenic changes in land use (Verschuren et al., 2000). Very severe droughts also afflicted the western United States during the MWP between 900 and 1300 AD (Cook et al., 2004).

ENSO forcing may be apparent in a five-year cycle that persisted through periods of high and low Nile flow (De Putter et al., 1998). Fraedrich et al. (1997) discerned eight almost synchronous abrupt changes in minimum and maximum Nile floods, many of which were associated with thirty-five- to forty-five-year persistence time scales. The 75.9 year periodicity identified by De Putter et al. (1998) from records of high flood levels during 950–1250 AD may be similar to the roughly 90 year periodicity evident in the Lake Lisan laminations from the Dead Sea Rift (Prasad et al., 2004). The increase in the frequency of high Nile floods after around 1250 AD may itself be linked to the increase in the south-west Asian monsoon inferred from proxy records of *Globigerina bulloides* collected from box cores from the Arabian Sea (Anderson et al., 2002). The reasons behind these periodicities remain obscure but appear to reflect changes in global atmospheric circulation linked to solar and other external controls (e.g., volcanic aerosol forcing) that were modulated by changes in SST in the equatorial Pacific and were reflected in ENSO events.

23.5 Volcanic eruptions and droughts

Using tree ring data from more than 300 sites across Asia, Cook et al. (2010) have established a precise chronology of very severe droughts and monsoon failure in that region. Four of the worst droughts were those of 1638–1641, 1756–1768, 1790/1792–1796 and 1876–1877. There are two matters of particular interest concerning the timing of these droughts. The first is that these severe droughts were also synchronous with wetter phases in other parts of Asia, so that there was considerable spatial variability across Asia (Cook et al., 2010; see also the commentary by Wahl and Morrill, 2010). The second matter, not discussed by Cook et al. (2010), is that each of these major drought intervals coincides very broadly with times when the volcanic dust

veil index (DVI) was high in the northern middle and high latitudes (Lamb, 1973). This latter point deserves further explanation. In a series of ground-breaking studies, Lamb compared the statistical relations between historic volcanic eruptions of well-constrained magnitude with any unusual changes in atmospheric circulation, near-surface temperature and net radiation flux (Lamb, 1970; Lamb, 1972, pp. 410–435; Lamb, 1977). In the process of doing this, he devised a series of volcanic dust veil indices (DVI) since 1500 AD, with the 1883 Krakatau eruption used as the standard DVI of 1000. The dust veil index of any explosive eruption can be estimated using a simple set of equations. For example, Lamb's DVI equation 3 is expressed as:

$$\text{DVI} = 4.4 \cdot q \cdot E_{\max} \cdot t \quad (23.1)$$

In this equation, q is the estimated volume in km^3 of solid matter dispersed as dust in the atmosphere, $E_{\max}$ is the greatest proportion of the earth at some time covered by the dust veil (taken as 1 for eruptions between 20°N and 20°S) and t is the total time in months elapsed after the eruption to the last readily observed effects in middle latitudes.

In regard to potential impact, only the finest ash particles (size range 0.5 to $2 \mu\text{m}$) are likely to persist in the stratosphere for periods of several or more years (Lamb, 1972, p. 411). The coarser particles (such as the 30 – $50 \mu\text{m}$ YTT deposited across India) will be rapidly removed from the atmosphere, in accordance with Stokes' Law (see Chapter 9, Equation 9.1). Niemeier et al. (2009) considered that the residence time of even fine ash in the stratosphere was more likely to be limited to months than years, so for the effects to persist for decades, some strong positive feedback processes and/or an interlude of more frequent global eruptions are needed (Lamb, 1970; Lamb, 1972).

A further impact is the direct change in albedo as a result of the deposition of a reflective ash mantle, leading to local near-surface cooling (Jones et al., 2007). Albedo, or reflectivity, is a measure of the amount of incoming solar radiation reflected back into the atmosphere. Williams et al. (2010a) noted that the colour of fresh YTT ash in India is white to very pale grey, which would imply an initial albedo value of 0.25 – 0.45 for the original ash mantle. Because the albedo under deciduous forest is 0.15 – 0.20 , there may have been a 5 – 30 per cent increase in the reflection of incoming solar radiation immediately after deposition of the YTT, leading to cooler surface temperatures, reduced mesoscale convective instability and less convective rain. However, this effect is unlikely to have persisted for more than a few years (Williams et al., 2010a). Future work will need to examine the processes involved in the association between major inputs of volcanic dust into the upper atmosphere and the weakening of the Asian summer monsoon.

Less well documented is the impact of eruptions on sea surface temperatures and on sea level fluctuations (Cazenave, 2005), with the ocean cooling persisting in some cases for decades (Gleckler et al., 2006). In this context, it is interesting to note that

Trenberth and Dai (2007) recorded several decades of drought in south-east Asia following the June 1991 Pinatubo eruption in the Philippines. They attributed the drought to a regional weakening of the hydrological cycle caused by cooling that had been engendered by the eruption. A growing body of evidence also now suggests a causal relationship between historic eruptions and El Niño events, resulting in sea surface temperature anomalies of near global extent (Adams et al., 2003; de Silva, 2003). Droughts are strongly controlled by changes in sea surface temperature (Lamb and Pepler, 1992; Williams and Balling, 1996), so another possible mechanism for prolonging the initial reduction in precipitation following an eruption (Parker et al., 1996; Trenberth and Dai, 2007) is the cooling of the ocean surface, which can last for many decades following the initial eruption (Gleckler et al., 2006).

Although low latitude volcanic eruptions appear to propagate more widely across the globe, as with the historic eruptions of Tambora (1815), Krakatau (1883), Agung (1963) and Pinatubo (1991), high latitude eruptions can also have a major impact on the tropics. For example, the 1783–1784 Laki volcanic eruption in Iceland was associated with significant cooling (-1 to -3°C) in the Northern Hemisphere during the boreal summer of 1783, weakening of the African and Indian monsoon circulation, reduced rainfall in Africa south of the Sahara and much-reduced Nile flow (Oman et al., 2006). Likewise, the 1982 eruption of El Chichón was in part responsible for severe drought in the Sahel. Oman et al. (2006) used the record of Nile flow to support a date of 939 for the eruption of Eldgjá volcano in Iceland, which was the largest high-latitude volcanic eruption of the last 1,500 years.

23.6 The 1968–1973 drought in the Sahel and Ethiopia

The *Sahel* (Arabic for ‘shore’ or ‘border’) refers to the semi-arid margins of the southern Sahara, and after two decades of above-average rains, it experienced a severe and prolonged drought starting in the late 1960s (Figure 23.8). Dwindling waterholes, pasture depletion and dying herds forced many nomadic pastoralists living in this region to move south in a desperate search for food and water for their animals and for themselves. The impact of erratic rains on a nomadic Tuareg community in central Niger, visited by the author in December 1974, is ably described by Bernus (1974) and is forcing a progressive move towards a more sedentary way of life (Grove, 1993). Elderly pastoral nomads and peasant farmers could still recall two earlier droughts of comparable severity to the 1968–1973 drought during the last century (Laya, 1975; Salifou, 1975), and historical records indicate that aridity became more pronounced in the Sahel at the end of the ‘Little Ice Age’ of Europe more than a century ago (Nicholson, 1976; Nicholson, 1978; Nicholson, 1980; Nicholson et al., 2012). Although the precise limits of the Little Ice Age are hard to define, the overall period lasted from about 1450 to 1850 AD, with significant glacier advances from 1550 to 1700 AD and again during the 1830s and 1850s (Lamb, 1977). Five cores

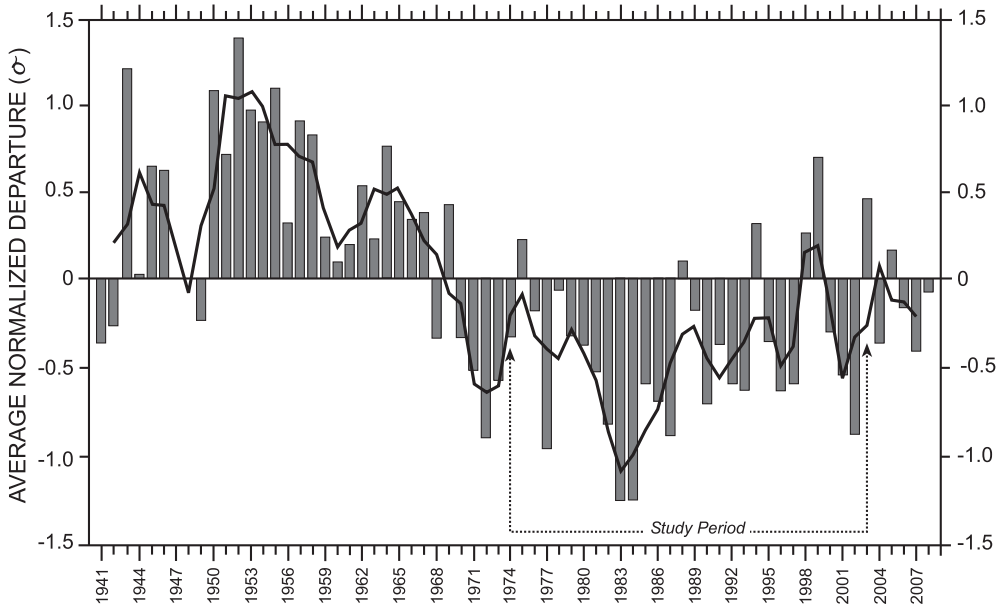


Figure 23.8. Time series (1941–2008) of average normalized April–October rainfall departure (σ) for twenty stations in the West African Sudan-Sahelian zone (11°–18°N) west of 10°E. (After Lélé and Lamb, 2010.)

from Lake Edward in Uganda indicated prolonged drought during the Little Ice Age in central Africa, in contrast to wet conditions over Lake Naivasha in the Kenya Rift Valley (Russell and Johnson, 2007). Differential migration of the ITCZ and shifts in the position of the ENSO system at this time could account for these differences. The ultimate causes of the drought that became severe in the late 1960s were immediately investigated (Dorize, 1974; Roche et al., 1975; Dorize, 1976; Pédelaborde, 1976) and are the subject of continuing study (Bell and Lamb, 2006; Slingo et al., 2008; Tarhule et al., 2009; Lélé and Lamb, 2010), including in the Horn of Africa (Segele et al., 2009a; Segele et al., 2009b; Segele et al., 2009c). An interesting attempt to place the incidence of droughts within a time scale longer than a century was conducted by Maley, whose pollen and limnological studies in Chad showed that high levels of Lake Chad coincided with lowered temperatures in Europe at least for the past millennium (Maley, 1973; Maley, 1981). Maley's concern with the possible influence of moist tropical air on the late Pleistocene growth of ice caps in the Northern Hemisphere (Maley, 1976) may seem a far cry from present-day droughts, but two well-argued reviews by Rognon (1974; 1976a) of Saharan and global paleoclimates suggest that the links may be there for the seeking.

One consequence of the Sahel drought was renewed interest in the nature and causes of rainfall variability in both the Sahel-Sudan zone and the Horn of Africa. Bell and Lamb (2006, p. 5343) described the severe Sahel-Sudan drought that began in

1968 as ‘among the most undisputed and largest regional climate changes experienced on the earth during the last half-century’. They noted that initial drought in 1968 was followed by pulses of severe drought in 1971–1973, 1977, 1982–1984 and 1987, with only modest rainfall recovery thereafter, followed by severe drought again in 2005. Most of the rainfall in this region occurs during June–September and comes from westward-propagating mesoscale convective systems, or ‘disturbance lines’. The marked decline in seasonal rainfall totals from the early 1950s to the mid-1980s (Figure 23.8) was due to a decrease in the size, organisation and intensity of these disturbance lines (Bell and Lamb, 2006). Another issue was whether or not rainfall in this region varied with the latitude reached by the Intertropical Front (ITF) during its seasonal northward migration. Lélé and Lamb (2010) found that peak rainfall occurred about 400 km south of the ITF. They also noted that meningitis epidemics occur in the dry conditions north of the ITF, while the incidence of malaria was confined to the warm, wet conditions south of the ITF. Sporadic locust outbreaks accompanied the southward retreat of the ITF, devastating ripening crops.

In the Horn of Africa, which includes Ethiopia and Somalia, rainfall during the June–September wet season reflects the intensity of the lower tropospheric south-westerly air masses from the equatorial Atlantic (Segele et al., 2009a). Wetter-than-average years coincide with lower-than-average mean sea level pressure over peninsular Arabia and higher-than-average pressure over the Gulf of Guinea, and vice versa (Segele et al., 2009b).

23.7 The albedo model of drought in the Sahel region of West Africa

The prolonged drought that afflicted the Sahel region of Africa as well as Ethiopia from 1968 onwards (Figure 23.8) also led to renewed debate about the role of humans in accentuating normal climatic variability and helping prolong severe drought. A question that is frequently asked is whether or not human activities such as overgrazing in savanna areas can cause or aggravate regional droughts. Charney proposed what he termed a ‘biogeophysical’ model for the contemporary drought in the Sahel region of West Africa (Charney, 1975; Charney et al., 1975; Charney et al., 1977). An alternative name is the albedo model of drought.

Albedo is the proportion of incoming solar radiation reflected from the earth’s surface in the form of outgoing, long-wave terrestrial radiation, and it is lower on dark, moist and well-vegetated surfaces and higher on bare, sandy or sparsely vegetated surfaces. If albedo increases, the surface will become cooler. In essence, overgrazing reduces plant cover and increases the ground surface albedo, or reflectivity. A cooler surface will mean less convection and hence less instability in the lower atmosphere. The result will be a reduction in convective rain. Less rain will mean less plant growth, and thus a vicious spiral of degradation or desertification will ensue along the desert margins (Figure 23.9; Table 23.4). The opposite occurs when there is less

Table 23.4. *Assumptions, testable conclusions and model results of the biogeophysical feedback models of the Sahel drought proposed by Charney (1975) and Charney et al. (1975, 1977). (After Williams and Balling 1996, p. 33.)*

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-
1. Overgrazing reduces vegetation cover
 2. Reduced plant cover increases albedo
 3. Increased albedo decreases surface net radiation
 4. Decreased surface net radiation results in surface cooling
 5. Surface cooling promotes subsidence of air aloft
 6. Subsidence decreases convection and cloud formation
 7. Reduced convectonal instability leads to less precipitation
 8. Additional drying in the Sahel region leads to regional climatic desertification, which positively feeds back to 1
 9. Atmospheric general circulation models show that an albedo increase from 14% to 35% north of the Intertropical Convergence Zone (ITCZ) results in a southward shift of a few degrees in the ITCZ
 10. Rainfall in the Sahel region is thus decreased in the model by 40% during the rainy season
-
-

grazing pressure, resulting in a greater plant cover, lower albedo, warmer soil surface, greater convection and, hence, more convectonal rainfall (Figure 23.9; Table 23.4).

The albedo model explains neither the synchronous nature of droughts in both hemispheres nor why they begin and end at about the same time. Nor does it account for even more severe historic droughts in Africa that occurred well before population increase and overgrazing were evident. There is now overwhelming evidence that the dominant cause of drought in the Sahel is a change in heat transport within the oceans, reflected in sea surface temperature (SST) anomalies in the Pacific, Indian and Atlantic oceans (Lamb, 1978a; Lamb, 1978b; Newell and Hsiung, 1987; Rasmusson, 1987; Street-Perrott and Perrott, 1990; Lamb and Peppler, 1991; Lamb and Peppler, 1992). For example, when the SSTs in the equatorial Atlantic are warmer than average and the Atlantic SSTs north of the equator and to the west of West Africa are lower than average, rainfall is reduced across the Sahel (Lamb and Peppler, 1991; Lamb and Peppler, 1992). It now appears that the drought in the Sahel probably involves a combination of land use change and dust and carbon aerosols interacting with more regional atmospheric and oceanic circulation processes (Menon et al., 2002; Giannini et al., 2003; Prospero and Lamb, 2003; Zeng, 2003).

While it is certainly true that local changes in surface albedo can cause changes in diurnal wind regime, it thus seems unlikely that they have more than a local influence. Although changes in albedo are no longer thought to be responsible for causing regional droughts, a local effect at the scale of tens of square kilometres has been documented within semi-arid Australia, and appears to be able to influence both the strength and direction of local sand-moving winds (Tapper, 1991).

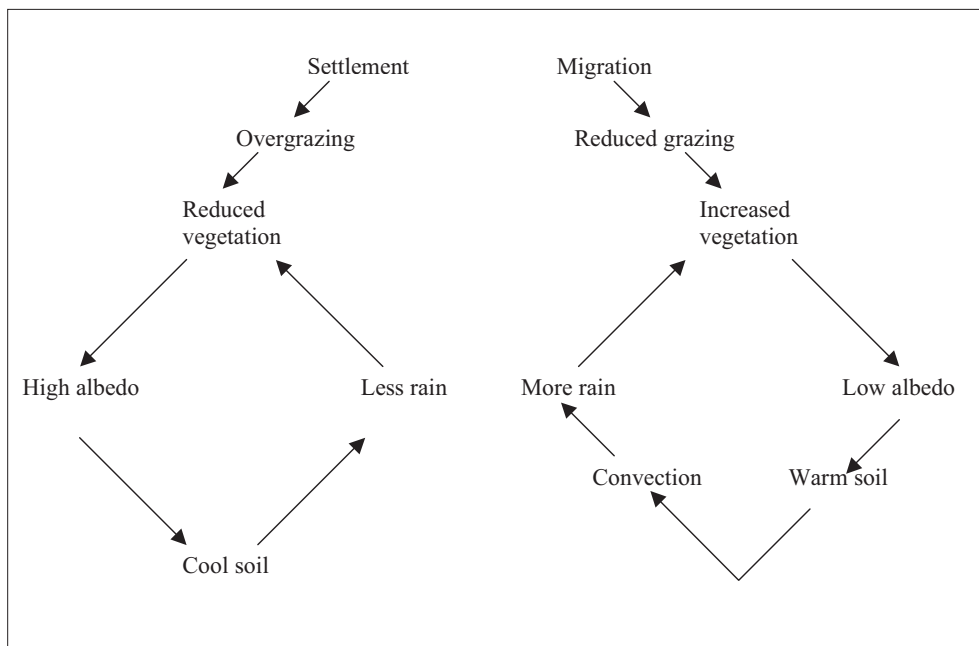


Figure 23.9. Hypothetical impact of overgrazing and reduced grazing, respectively, on plant cover, albedo and rainfall in drylands. (After Williams and Balling, 1996, fig. 2.3, adapted from Krebs and Coe, 1985, and Sinclair and Fryxell, 1985.)

However, widespread overgrazing can lead to the accelerated loss of top-soil from wind and water erosion, resulting in a breakdown in soil structure, lower infiltration capacity and much-reduced soil moisture storage, which causes widespread ‘edaphic drought’, the consequences of which we outline in [Chapter 24](#). There have been some ingenious attempts to model changes in run-off related to changes in land use. For example, Parida et al. (2006) made innovative use of Artificial Neural Network techniques to model run-off coefficients in a partly urbanised catchment in eastern Botswana. The results enabled the authors to distinguish between the impact of climatic factors on the one hand and land use changes on the other in controlling the increase in run-off coefficients between 1978 and 2000. This is a very important finding, for it enables the influence on run-off of different environmental variables to be quantified accurately and reasonably precisely, and will be of considerable value when applied to forecasting likely changes in other catchments in the wider region.

23.8 The North Atlantic Oscillation and drought

Forty years ago, Namias (1972) drew attention to the links between drought in north-east Brazil and atmospheric circulation in the Northern Hemisphere. In particular, he reported that the highly variable interannual rainfall in north-east Brazil depended

on the degree of cyclonic activity or of blocking in the Newfoundland-Greenland area during the Northern Hemisphere winter and spring, with blocking leading to severe drought in north-east Brazil. Later work showed a close link between SST in the South Atlantic and drought in north-east Brazil, with drought unlikely when the SSTs are warmer than 24.5°C and flooding unlikely when the SSTs are colder than 23.5°C (Markham and McLain, 1977). Chang et al. (1997) found a decadal oscillation between SST in the tropical Atlantic on either side of the equator, and concluded that this dipole mode reflected ocean-atmosphere interactions.

23.9 The Pacific Oscillation and drought

We have seen that historic floods and droughts are associated with the zonal Walker Circulation (see Section 23.2), which propagates from east to west across the equatorial Pacific and reflects the magnitude and frequency of ENSO events. The Pacific Oscillation is another contributor to climatic instability in this region (Salinger et al., 2001; Mantua and Hare, 2002; Pierce, 2002). Salinger et al. (2001) recognised three phases of what they termed the Interdecadal Pacific Oscillation (IPO) during the twentieth century: a positive phase (1922–1944), a negative phase (1946–1977) and another positive phase (1978–1998). During the most recent positive phase, mean sea level atmospheric pressure (SLP) in the area west of 170°W increased compared to the previous negative phase and decreased east of 170°W. The result was a 30 per cent increase in annual precipitation north-east of the South Pacific Convergence Zone. The IPO modulates the influence of ENSO in the Pacific region and can accentuate it in some areas while weakening it in others. ENSO events only last about a year and generally occur about every five to eight years, in contrast to IPO phases, which last for several decades. Using corals and tree rings, Mantua and Hare (2002) found evidence of periodicities associated with the Pacific Decadal Oscillation (PDO) dating back to 1600, and they identified two main periodicities, one of fifteen to twenty years and the other of fifty to seventy years.

23.10 Assessing drought severity

There have been various attempts to assess the severity of drought, of which the best-known and most widely used is the Palmer Drought Severity Index (PDSI) (Palmer, 1965). This index is a measure of drought severity and soil moisture deficit, and it was designed to allow a rapid comparison between different regions in the United States affected by drought, based on the rapid mapping of drought-affected areas. Cook et al. (1999) provided a detailed explanation of the methods used to reconstruct the PDSI for the continental United States between 1700 and 1978, based on 425 tree-ring chronologies from across the United States, and concluded that the 1930s ‘Dust Bowl’ drought was the most severe in this region since 1700. Cook et al. (2004; 2007)

used an expanded database and produced the North American Drought Atlas. A similar approach by Cook et al. (2010) resulted in the Monsoon Asia Drought Atlas, which was based on a network of 327 tree-ring chronologies. The results showed four intervals of severe and widespread drought during the last 1,000 years, which they named the 'Ming Dynasty drought' of 1638–1641, the 'Strange Parallels drought' of 1756–1768, the 'East India drought' of 1790–1796 and the 'late Victorian great drought' of 1876–1878. The PDSI is very useful for comparing the relative severity of different historic droughts for which there is a detailed tree-ring chronology; it is less useful for evaluating recent trends.

The PDSI calculates potential evaporation (E_{pot}) from temperature data using the empirical Thornthwaite equation (Thornthwaite, 1948). However, a number of factors besides temperature control E_{pot} , including wind speed, humidity and near-surface radiation. Despite early claims that historical droughts in the United States were on the increase, detailed analyses in the 1990s showed no evidence of an increase in drought frequency, intensity, magnitude or duration (Karl and Heim, 1990; Idso and Balling, 1992; Soulé, 1993). A more recent study by Sheffield et al. (2012) compared the PDSI based on the Thornthwaite E_{pot} equation (PDSI_Th) and the PDSI in which E_{pot} is estimated from the physically based Penman-Monteith equation (PDSI_PM). The results show that PDSI_Th overestimates the global increase in drought, whereas the PDSI_PM shows an increase in some areas and a decrease in others. These results cast doubt on the IPCC (2007a) conclusion that 'more intense and longer droughts have been observed over wider areas since the 1970s, particularly in the tropics and subtropics. Increased drying linked with higher temperatures and decreased precipitation has contributed to changes in drought'. One reason why the IPCC conclusion may be badly flawed concerns the use of temperature to assess E_{pot} using the PDSI_Th index. If global temperatures are indeed increasing, then use of a drought formula based on temperature will inevitably show an increase in drought frequency. Overall, Sheffield et al. (2012) found little evidence of change in global drought over the past sixty years, and they noted that severe droughts in the 1950s and 1960s took place well before the rapid increase in global warming of the last few decades. The 2012 IPCC report on extreme events is more circumspect in predicting drought trends, observing that there has been undue reliance on the PDSI and possible overestimation of the increase in regional and global droughts (Seneviratne et al., 2012).

23.11 Extreme events as partial analogues of early Holocene environments

Modern extreme events can sometimes provide insights into the nature of early Holocene prehistoric environments in ways that are simply not possible using global atmospheric circulation models. Let one example suffice. The year 1999 was unusually wet in many regions, with severe floods in Australia, China and India and very high Nile discharge. In central Sudan, the rainfall in July 1999 was unusually heavy and

persisted throughout August, causing severe floods in much of central Sudan, including Khartoum. Extreme rainfall events in central Sudan are historically associated with a warm equatorial Indian Ocean, a strong summer monsoon over both Africa and India, a northward shift of the Intertropical Convergence Zone earlier and further north than usual, and the presence of deep, well-developed westerly air masses accompanied by a strong Tropical Easterly Jet that allow more moisture transport into Africa from the South Atlantic via the Congo Basin, which leads to very heavy precipitation in the Ethiopian uplands and the central Sudan (Camberlin, 1997; Camberlin et al., 2001; Mo et al., 2001; Osman and Shamseldin, 2002; Vizzy and Cook, 2003). Satellite imagery revealed that the intense 1999 wet season rains in central Sudan filled the depressions between the extensive dunes immediately east of the lower White Nile and recreated the geography of early Holocene times (Williams and Nottage, 2006). This was a time when bands of Mesolithic hunter-gatherers used to camp seasonally on top of the dunes overlooking the flooded ponds between them in order to fish, harpoon hippos and collect *Pila wernei* shells for food (Adamson et al., 1974; Clark, 1989).

23.12 Conclusion

Changes in sea surface temperatures in the equatorial Pacific Ocean and adjacent Indian Ocean associated with the Southern Oscillation are a major source of inter-annual variations in the strength of the summer monsoon. The Southern Oscillation reflects the surface atmospheric pressure differences between the eastern and western limbs of the equatorial Pacific and gives rise to the Walker Circulation. The strength of the Southern Oscillation is measured as the pressure difference between Darwin and Tahiti, expressed as the Southern Oscillation Index (SOI). In years when the SOI is strongly negative (i.e., there is anomalously high pressure at Darwin and anomalously low pressure at Tahiti), there tends to be below-average precipitation or droughts in north-east Brazil, eastern Australia, eastern China, peninsular India, north and central Thailand, Java, parts of southern Africa and at the Ethiopian headwaters of the Blue Nile. Conversely, when the SOI is strongly positive, these same regions experience above-average precipitation and often severe floods.

These patterns have changed only slightly over the past hundred and more years, but there are signs of a change in the previously well-established, statistically determined correlations (or teleconnections). For example, since the 1970s, summer rainfall in India has become less linked to El Niño-Southern Oscillation (ENSO) events and more linked to phase changes in the Indian Ocean Dipole. There is also evidence that from 1977 to 2006, when tropical sea surface temperatures were the highest on record and mean surface atmospheric pressure at Darwin was the highest recorded, there was a weakening of the Walker Circulation, and June–December values of the SOI were the lowest on record. A corollary to these changes in previous patterns is that global

warming will probably lead to changes in the magnitude and frequency of ENSO events in ways that remain uncertain. Those areas now most vulnerable to interannual climatic variability will continue to be the arid and semi-arid areas, but the seasonally wet tropics and the dry subhumid regions that now receive 750–1,500 mm of rain a year will become increasingly sensitive to future changes in temperature, evaporation and precipitation. Biomass burning in the seasonally wet tropics of the Asia-Pacific region is severe during dry years and has the potential to change regional climate. There are some signs that the increase in carbon particles in the lower atmosphere arising from biomass burning and the use of fossil fuels is leading to increasing desiccation of northern China and increasing flooding in the south (Menon et al., 2002).

Volcanic eruptions can lead to short-term cooling and more prolonged drought, while changes in albedo caused by overgrazing can have a local effect on climate but are unlikely to be the main cause of regional droughts.

Desertification: causes, consequences and solutions

The wind grew stronger. The rain crust broke and the dust lifted up out of the fields and drove gray plumes into the air like sluggish smoke. . . . The finest dust did not settle back to earth now, but disappeared into the darkening sky.

All day the dust sifted down from the sky, and the next day it sifted down. An even blanket covered the earth. It settled on the corn, piled up on top of fence posts, piled up on the wires; it settled on roofs, blanketed the weeds and trees.

John Steinbeck (1902–1968)

The Grapes of Wrath (1939)

24.1 Introduction

In the quotation cited at the head of [Chapter 23](#), Aldo Leopold (1949) commented that: ‘The effort to control the health of land has not been very successful. It is now generally understood that when soil loses fertility, or washes away faster than it forms, and when water systems exhibit abnormal floods and shortages, the land is sick’. He was referring very clearly to what we now call desertification. He went on to argue that ‘the fallacy the economic determinists have tied around our collective neck, and which we now need to cast off, is the belief that economics determines **all** land use’.

Desertification is an elusive concept and has proved very hard to define. The processes leading to desertification have been widely debated, often without satisfactory resolution. One reason for this lack of accord has been a polarisation of the debate, with one side claiming human activities as *the* cause of desertification and the other claiming that climate was the primary cause of desert expansion into hitherto unaffected areas. In order to understand the reasons behind the international adoption of the current definition, it will be useful to examine the history of this debate (Grove, [1974](#); Darkoh, [1989](#); Warren and Khogali, [1992](#); Mainguet, [1994](#); Thomas and Middleton,

1994; Kassas, 1995a; Williams and Balling, 1996; Williams, 2000; Williams, 2002b; UNEP 2002a; UNEP, 2002b; Imeson, 2012).

Desertification was for some years defined by the United Nations Environment Programme (UNEP) as ‘land degradation in arid, semi-arid and dry subhumid areas resulting *mainly from adverse human impact*’ (UNEP 1992a, italics added), but since June 1992 it has been redefined as ‘land degradation in arid, semi-arid and dry subhumid areas resulting from *various factors including climatic variations and human activities*’ (UNEP, 1992b, italics added). This definition of the climatic areas involved excludes the hyper-arid regions such as the Sahara and the Atacama deserts, but it does comprise about 40 per cent of the land area of the globe, including significant parts of Eurasia, Africa, Australia and the Americas.

The currently accepted definition was the one agreed upon at the United Nations Conference on Environment and Development (UNCED), which was held in Rio de Janeiro in June 1992 and popularly known as the Earth Summit (UNCED, 1992). In contrast to the superseded UNEP definition, which stressed human mismanagement and ignored any influence of climatic variations, the 1992 UNCED definition took into account climatic fluctuations, especially prolonged droughts. The main reason for this change in perception was the impact of the severe 1968–1974 drought in the Sahel region, which first became apparent along the south-western margins of the Sahara and spread progressively eastwards to Ethiopia and Somalia, causing many tens of thousands of deaths from starvation when the harvests from rain-fed agriculture failed. (*Sahel* is an Arabic term for ‘shore’, and it refers geographically to the region bordering the southern Sahara, including countries such as Mali, Niger and Chad.) This prolonged drought, which was to persist with only minor interruptions for nearly thirty years (Chapter 23), stimulated renewed scientific interest in the causes of severe droughts and revealed very starkly how a sustained lack of rain can contribute to widespread land degradation. At the International Geophysical Conference in Grenoble in August 1975, French hydrologists and meteorologists drew attention to the impact of this drought. Two years later, growing concern over water shortages and land degradation prompted the United Nations to host a Water Conference in Argentina in March 1977.

Five months later, UNEP organised a United Nations Conference to Combat Desertification at the UNEP headquarters in Nairobi, at which a number of attempts were made to define the elusive concept of ‘desertization’, or the spread of desert-like conditions (United Nations, 1977; United Nations, 1978; Glantz, 1987; Williams, 2002b). One suggestion was to restrict the meaning to denote the extension of such conditions into semi-arid areas only, defined as areas in receipt of 50–300 mm of rain a year. Others argued that areas of higher rainfall should be included, a view endorsed by UNEP, so ‘desertization’ was defined as ‘the spread of desert-like conditions in arid or semi-arid areas up to 600 mm, due to man’s influence or climatic change’ (United Nations, 1977; Glantz, 1987). Later workers ignored the climatic factor, and some

even argued that drought was a human construct arising from destructive forms of land use – an odd way to ignore the reality of prolonged below-average rainfall.

As time passed, the UNEP definition stressing ‘adverse human impact’ was widely adopted, to the dismay of many African nations, who were all too aware of the adverse impact of droughts and so were able to lobby effectively at the 1992 Rio Earth Summit for a change in definition and for the creation of a future international convention to deal with desertification. The African environment ministers had become frustrated by the lack of any substantive progress in dealing with desertification in Africa and therefore met in November 1991 in Abidjan in order to prepare a common position paper for Africa to present at the 1992 Earth Summit (Kassas, 1995b; Williams, 2002b). They were unanimous in requesting that a new convention to combat desertification be a specific recommendation for inclusion in *Agenda 21*, and with the backing of many other countries who were facing similar problems, including China (China’s *Agenda 21*, 1994; Ci Longjun, 1998), they were successful. Chapter 12 of *Agenda 21* is called ‘Managing fragile ecosystems: combating desertification and drought’ and is a lucid analysis of the causes and consequences of desertification, with some very practical action plans for use at national and international scales (UNCED, 1992).

The great African droughts of the 1930s and the 1970s triggered considerable speculation as to the respective role of climatic fluctuations and human influences on land degradation along the margins of the Sahara. Many observers blamed the destruction of the vegetation and the mobilisation of hitherto fixed and stable dunes entirely on human mismanagement in the form of overgrazing and the clearing of forest and woodland for arable land. Some even argued that the Sahara Desert was itself at least in part a product of adverse human impact. Phrases like ‘desert encroachment’ or ‘the advancing desert’ were bandied about with more enthusiasm than accuracy (see also Grove, 1974, for an incisive and elegant critique of such views in regard to Africa). Such sensationalism took no account of the long geological history of the deserts, nor of the fact that the deserts long predate human origins (see Chapters 3, 17 and 18). While the deserts may have helped shape the prehistoric cultures of the occasional desert dwellers, there is scant evidence that the reverse applies, at least until more recent times.

Political and scientific awareness of desertification was enhanced by the publication of two quite different editions of the UNEP World Atlas of Desertification (UNEP, 1992a; UNEP, 1997). One of the great merits of both atlases lies in the detailed case studies prepared by local specialists drawn from throughout the desert world, which diagnose the local and regional causes of land degradation and identify a variety of solutions. In the five years between the publication of the first and second editions of these two very useful atlases, there was already a much clearer scientific appreciation of the nature and causes of climatic variations, including global floods and droughts. This is very clearly reflected in the thoughtful editorial discussion of climatic fluctuations by Middleton and Thomas in the second edition (UNEP, 1997).

It is also worth noting that a number of workers reject the very notion of desertification as defined by international organisations such as UNEP, arguing that local farmers are far more in tune with environmental fluctuations than outside agencies are willing to acknowledge and that they show a high degree of resilience and a considerable capacity to adapt to change (Bassett and Crummey, 2003).

In addition, there is good evidence in support of ‘natural desertification’, that is, land degradation caused by a sequence of geomorphic processes initiated, directly and indirectly, by long-term changes in climate that have nothing to do with human activities (Avni, 2005; Avni et al., 2006; Avni et al., 2010). We will consider these studies first before evaluating the impact of human activities on the landscapes of deserts and their margins. The manifestations of dryland degradation or desertification are as varied as the individual causes (Table 24.1), indicating that it would be naive to seek a single cause. A potentially useful approach would be to examine a number of examples drawn from different parts of the desert world in order to learn from history and see if we can identify any general prerequisites for achieving optimum land use in arid lands. This chapter seeks to do that.

24.2 Natural desertification

Gully erosion has been widely considered as one of the more obvious signs of accelerated soil loss brought about by human mismanagement, and it is not hard to find apparently convincing evidence of such processes in many parts of the desert world. However, a degree of caution is advisable before attributing the cause of gully erosion to human actions. In a number of instances, the gullies were already active along the valley floors well before humans had occupied those valleys. Yoav Avni and his co-workers monitored a number of gullied valleys in the semi-arid Chifeng region of Inner Mongolia (northern China) and in the Negev Highlands of southern Israel (Avni, 2005; Avni et al., 2006; Avni et al., 2010). They found that initial deposition of loess (wind-blown desert dust, see Chapter 9) across the hills and valley sides during the late Pleistocene (between about 70 and 15 ka) was followed by erosion and redeposition of the valley-side loess mantles and their subsequent accumulation in alluvial fans and valley fills. A change in rainfall intensity and a reduction in dust influx during the end of the Pleistocene triggered a change from reworked loess accumulation along the valley floors to incipient vertical erosion and gully incision. Thereafter, the gully network extended up-valley through a combination of headwall erosion and some lateral erosion from bank collapse, leading to a loss of potentially valuable soil for cultivation and plant cover for grazing. The key factor here was the progressive exposure of bedrock surfaces along the valley sides and the ensuing increase in run-off and its capture by the expanding gully system.

In one instance in the Negev, two valleys of similar size and geology were compared, since one had been farmed in medieval times more than a thousand years ago and the

Table 24.1. *Possible causes and consequences of desertification. (From Darkoh, 1989; Williams et al., 1995; Darkoh, 1999; Williams, 2002b.)*

Causal factor	Consequences
<i>Direct land use</i>	<i>Physical processes affected</i>
Over-cultivation (shorter fallows; mechanised farming)	Decline in soil structure and soil permeability; depletion of soil nutrients and soil organic matter; increased susceptibility to erosion; compaction of soil; sand dune mobilisation
Overgrazing	Loss of biodiversity and biomass; increased wind and water erosion; soil compaction from trampling; increased run-off; sand dune mobilisation
Mismanagement of irrigated lands	Waterlogging and salinization of soil; lower crop yields; possible sedimentation of water reservoirs
Deforestation (burning to clear land; fuel and fodder collection)	Promotes artificial establishment of savanna vegetation; loss of soil-stabilising vegetation; exposed and eroded soil increases soil aridity; more frequent dust storms; sand dune mobilisation
Exclusion of fire	Promotes growth of unpalatable woody shrubs at the expense of herbage
<i>Indirect government policies</i>	<i>May cause over-exploitative land use practices</i>
Failed population planning policies	Increases need for food cultivation, and so may lead to over-exploitation of marginal land
Irrigation subsidies	Exacerbates flooding and salinization
Settlement policies/land tenure	Forces settlement of nomads; promotes concentrated use of land, which often exceeds the local carrying capacity
Improved infrastructure (e.g., roads; large-scale dams; canals; boreholes)	Although often beneficial, can aggravate the problem by attracting increased livestock and human populations or increasing risk from salinization; possible lowering of groundwater-table below dams and around boreholes; silting up of reservoirs; waterlogging; promotes large-scale commercial activity with little local benefit; flooding may displace people and perpetuate cycles of poverty
Promotion of cash crops and push towards national and international markets	Displaces subsistence cropping; pushes local people into marginal areas to survive; promotes less resilient monocultures; fosters expansion and intensification of land use
Price increases on agricultural produce	Incentive to crop on marginal lands
High interest rates	Forcing grazing or cultivation to levels beyond land capacity
War	Valuable resources, both human and financial, are expended on war at the expense of environmental management and the needs of the people; large-scale migration with resultant increased pressure on receiving areas
<i>Natural</i>	
Extreme drought	Decreased vegetation cover and land more vulnerable to soil erosion; creates an environment that exacerbates over-exploitation
Extreme flooding	Loss of arable land, houses and infrastructure; displacement of people; increased land use pressure on receiving areas
Ecological fragility	Impact of land use practices (will also depend on resilience of environment)

other had not (Avni et al., 2006). The stone terraces constructed by the early Byzantine farmers in the cultivated valley had helped trap several metres of colluvial-alluvial loess and gravel, and where these structures had survived undamaged, those parts of the valley were still cultivated by the local Bedouin.

The methodology used in these carefully devised studies is relatively straightforward and could profitably be applied more widely. The first step is to date the valley fills using a combination of radiocarbon and luminescence dating methods allied with detailed mapping and logging of the main landforms and stratigraphic units. This allows the timing of the change from a regime dominated by loess and alluvial loess deposition to one of erosion to be determined reasonably accurately and compared with the archaeological evidence of human settlement in the region. A further by-product of this approach is that it allows rates of sedimentation to be determined for different intervals of time, together with quantitative estimates of rates of gully headwall retreat.

The conclusion from each of the three studies was that land degradation caused by gully erosion is indeed contributing to desertification but was not in any way brought about by human actions. On the contrary, well-designed and properly maintained soil and water conservation measures helped arrest the process in part of one valley for more than a thousand years, indicating that human actions can be both beneficial and enduring.

One unexpected cause of gully erosion that had nothing to do with either climate or humans was an earthquake that led to the formation of a deep fissure in the welded tuffs and overlying sediments on the floor of one of the calderas within the K'one volcanic complex near the eastern margin of the Ethiopian Rift Valley where it begins to merge into the southern Afar Rift (Williams, 2003). The earthquake occurred some 5,000 years ago, and the gully erosion that followed created a veritable 'badland' topography incised into horizontally stratified sediments that contain Middle Stone Age obsidian artefacts in the lower units and Late Stone Age and Neolithic artefacts and charcoal (dated by ^{14}C) in the uppermost units. Although the government of the day was inclined to blame the local charcoal burners for the gully erosion, they were unaware that the gullies were no longer active and had probably been stable for several thousand years, as shown by a detailed plane table survey of the entire gully complex (Williams, 2003).

The vegetated dunes between Tahoua and Abalak, located about midway between Niamey and Agades in central Niger provide another instance of gullies formed during extreme events (Talbot and Williams, 1978; Talbot and Williams, 1979). During the course of a camel survey of the Wadi Azouak Basin in late 1974, near the end of a long interval of drought, they observed a series of sandy alluvial fans located on the lower slopes of some fixed dunes near Janjari village. Older fan sediments were exposed in the banks of incised channels upstream of the most recent fan, which had formed during an intense rainfall event in the previous wet season earlier in 1974. Using the

most recent fans as a guide, they estimated that the impact of this rainfall event was confined to an area no more than about 10–15 km in radius. The downpour caused channel entrenchment above the fan apices and fan sedimentation below. Buried soils exposed in the banks of the most recent gully showed repeated episodes of dune dissection, fan deposition and soil development. The soils would have formed when the fan surface was once more vegetated and stable. The youngest soil probably formed during a regionally wetter interval in the Sahel dated to about 150–350 years ago. Their overall conclusion was that the 1974 phase of dune dissection was no greater in scale than previous Holocene phases and that there was every reason to assume that recovery would again be possible. Needless to say, this conclusion would probably apply to other parts of the Sahel, provided that grazing allowed re-establishment of the plant cover.

24.3 Detecting desertification using changes in plant cover

One unresolved problem in detecting a climatic influence on desertification concerns the high degree of variability in the annual plant cover along the desert margins. Tucker et al. (1991) used remote sensing data for the 1980–1990 period in an attempt to plot the expansion and contraction of the Sahara Desert during that decade and found that there was such a high degree of variation in plant cover relative to both longitude and latitude that they were forced to conclude that ten years were too few for any meaningful trend to be discernible. The previous year, Nicholson et al. (1990) had sought to use the Normalized Difference Vegetation Index (NDVI) to evaluate the response to rainfall in the Sahel and East Africa. They discovered that there was a linear correlation between NDVI and mean precipitation.

Tucker and Nicholson (1999) revisited this problem nearly a decade later and claimed that the longer period of observation did enable them to detect significant trends in the response of plant cover to decadal variations in rainfall. This type of research is needed in all desert margin systems, given that the response of the NDVI to annual rainfall may well vary from region to region, depending on the different soil types and plant associations. For example, acacia species in the Sudan growing on sandy soils require up to 35 per cent less precipitation than the same species in the same latitude growing on clay soils (Smith, 1949; see also [Chapter 4](#)).

One example will suffice to show how past estimates of changes in plant cover along the southern edge of the Sahara have been taken out of context and misused. From 21 October to 10 November 1975, the ecologist Dr Hugh Lamprey completed an aerial survey of the plant cover along the desert margin of northern Sudan (Lamprey, 1975). A comparison with the northern limit of semi-desert vegetation in the Sudan mapped by Harrison and Jackson (1958) revealed a change for the worse, prompting Lamprey to conclude that ‘the desert boundary has shifted southward by an average of about 90–100 kilometres in the last 17 years’. Given the high degree of interannual variability in

the plant cover along the southern borders of the Sahara, such an averaging has very little meaning. The 1950s were wetter-than-average years in northern Sudan, and the Lamprey survey came after a decade of below-average rain.

Other workers have made similar poorly supported claims, generally in the wake of several years of drought. See, for example, the publications of the forester E.P. Stebbing (1935; 1937a; 1937b; 1938), with their apocalyptic titles about ‘the encroaching Sahara’ and ‘the man-made desert in Africa’. Unfortunately, this type of strident oversimplification becomes widely circulated and quoted out of context. For instance, in their popular account *Population, Resources, Ecology: Issues in Human Ecology*, Paul and Anne Ehrlich write expansively that ‘the vast Sahara desert itself is largely man-made, the result of overgrazing, faulty irrigation, deforestation, perhaps combined with a shift in the course of a jet stream. Today the Sahara is advancing southward on a broad front at a rate of several miles per year’ (Ehrlich and Ehrlich, 1970, p. 166). Such a claim (echoing Lamprey, 1975) is patent nonsense. We have already seen in Chapters 1 to 3 that the deserts predate the advent of humans by many millions of years and that they are where they are for sound geographical reasons that have nothing to do with humans. What we can say with some confidence is that during colder, drier and windier intervals throughout the late Quaternary, such as during Last Glacial Maximum 20,000 years ago, many of the tropical deserts of Africa and Asia were indeed more extensive than they are today, meaning that the presently vegetated and stable dunes along their now semi-arid margins were active at those times, as was discussed in Chapter 8.

24.4 Causes of anthropogenic desertification

As an exercise in perception, the author once asked senior government natural resource managers from many parts of Asia and some parts of Africa and South America to list what they considered to be the primary causes of desertification. The resulting list of causes tallies almost exactly with those identified in the chapter on desertification in *Agenda 21*, which had been published by that time and widely circulated among relevant government departments and agencies. The causes identified were deforestation, overgrazing, population pressure, inappropriate agricultural practices (shifting cultivation with too short a fallow; cultivating on very steep slopes; overuse of fertilizers), over-irrigation, poor drainage, fire, drought, floods and climate change. Conversations that the author has held at intervals over the last fifty years with cultivators and pastoralists in northern Africa (Sudan, Ethiopia, Niger, Kenya, Somalia, Tunisia, Algeria), arid northern China (Xinjiang Province and Inner Mongolia) and the drier parts of India (Rajasthan, Andhra Pradesh and Madhya Pradesh) suggest that other causes may be of equal importance, at least at a local or regional level. These include low prices for cash crops and high prices for subsistence items such as grain, invasions of weeds and plants unpalatable for livestock, an increase in the frequency

of extreme events such as droughts, an influx of immigrants and livestock from other regions, and restricted access to former grazing lands. Not surprisingly, local perceptions and government perceptions did not always coincide, and sometimes they were radically different (see also [Table 24.1](#)).

Williams and Balling (1996) provided a comprehensive account of the interactions between desertification and climate, and the two UNEP Atlases of Desertification included a number of useful case studies from across the desert world (UNEP 1992a; UNEP, 1997). The direct consequences of desertification are now reasonably well-known. However, in their early stages, they may be hard to discern, except by prescient and experienced observers, as was the case with the Aral Sea (Glantz, 1999). The causes are far more complex and involve a web of interlinked social, economic and environmental factors, as illustrated in [Figures 24.1](#) and [24.2](#).

24.5 Interactions between biophysical, economic and social factors

[Figure 24.1](#) represents a higher level of complexity than the simple biogeophysical or albedo model reviewed in [Chapter 23](#) and illustrates some of the interactions at global and regional scales between physical, chemical and biological processes involved in land degradation in drylands. The importance of considering how feedback mechanisms operate in different ways at different spatial scales and institutional levels is illustrated in [Figure 24.2](#). One matter that has become increasingly evident over the past few decades concerns failures of past attempts to control desertification through a lack of attention to social and economic factors at local and regional scales. Bassett and Crummey (2003) provide a wealth of examples from Africa that underline the importance of taking into account and indeed being willing to learn from local knowledge and experience. Many attempts to ‘combat’ desertification have failed because they were imposed from above and took no notice of local customs and capacities. One example concerns the restriction imposed by governments on traditional grazing rights, which leads to overgrazing, for which the pastoral nomads are then blamed. The creation of National Parks without consulting of those most affected – the people who were previously living in the parks or were accustomed to moving through them with their herds – has in a number of instances led to anger and violence, which is not a good recipe for long term land management (Western, 2002). Conservation measures will seldom work until fundamental issues of poverty alleviation, food security and health are also dealt with effectively. We return to this question in the final chapter of this volume ([Chapter 26](#)).

24.6 Consequences of desertification

In a pioneering study of the global context of desertification in Australia, Mabbutt (1978) defined desertification as ‘a change to more desertic conditions’, which of

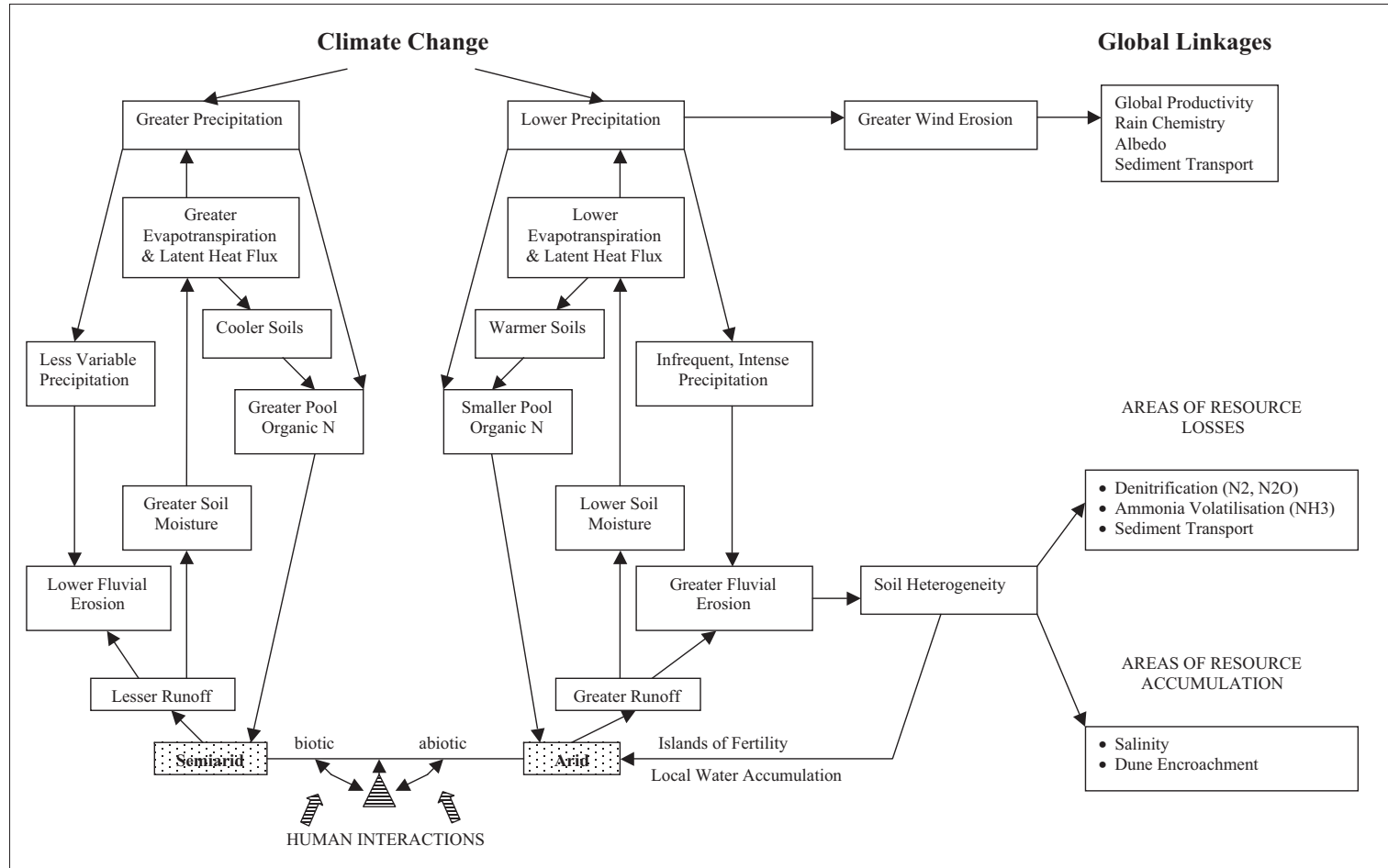


Figure 24.1. Linkages between global changes, human activities and desertification. (Redrawn from Williams and Balling, 1996, fig. 1.4.)

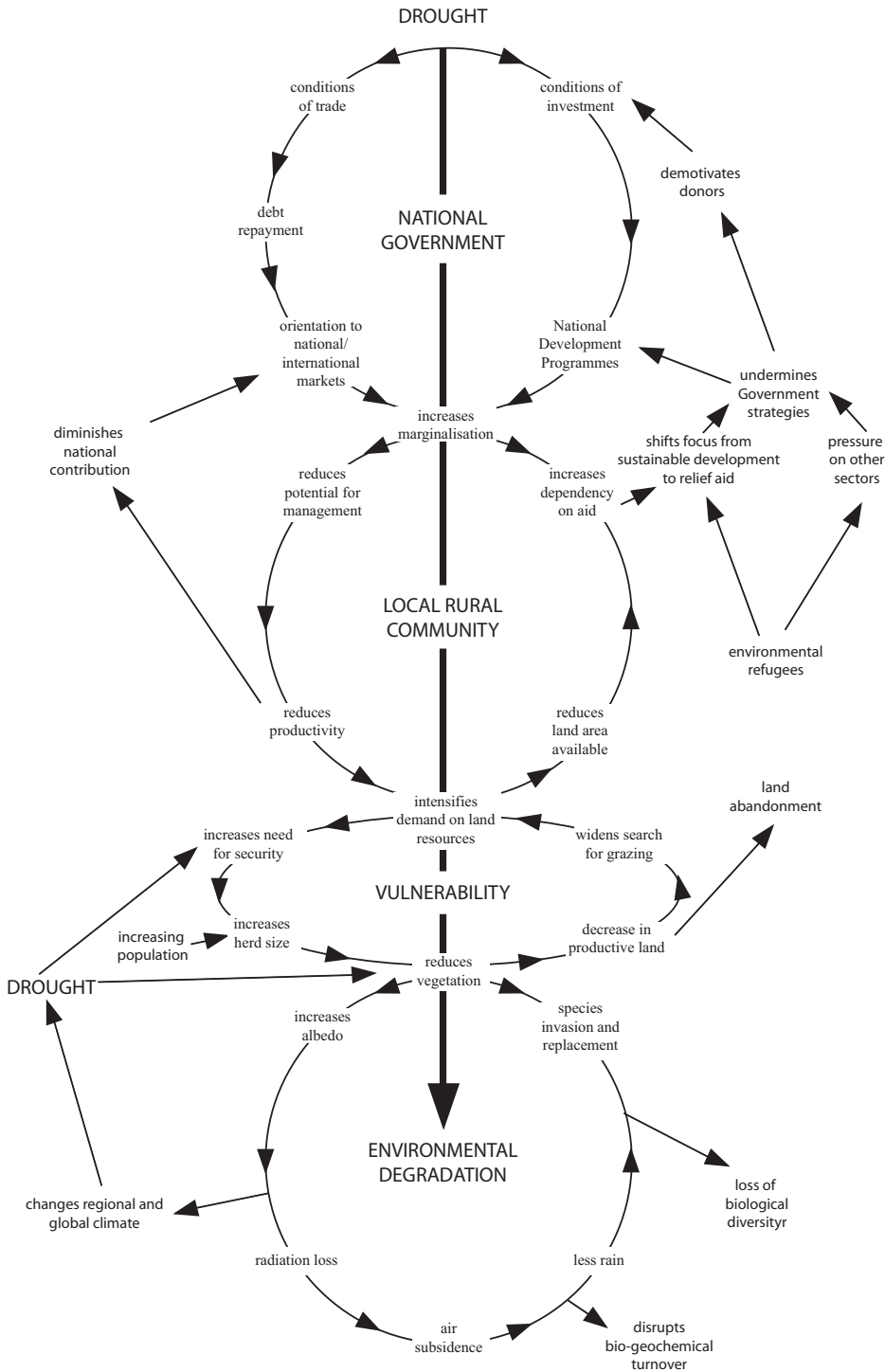


Figure 24.2. Examples of self-reinforcing mechanisms (positive feedbacks) at international, national and local levels resulting in rangeland desertification. (Redrawn from Williams and Balling, 1996, fig. 6.1.)

course begs the question of just what ‘desertic’ means. However, and more usefully, he also pointed out that desertification caused impoverishment of ecosystems, accelerated soil degradation, reduced plant and animal productivity, and impoverishment of dependent human livelihood systems. When there was a combination of climatic stress and land use pressure, the almost inevitable result was land degradation and, in extreme cases, famine (Mabbutt, 1978). Other workers have also emphasised the outward signs of desertification in affected landscapes (Kassas, 1995a; Williams and Balling, 1996; Williams, 2000a; Williams, 2002b; Williams, 2003). The causes of desertification will vary from region to region, as discussed in Sections 24.7 to 24.10. When there is a combination of climatic stress and land use pressure, the almost inevitable result is desertification. Some of the more immediate signs of desertification may include any or all of the following:

- Accelerated erosion by wind and water;
- A decline in soil structural stability, with an ensuing increase in surface crusting and surface run-off, and a reduction in soil infiltration capacity and soil moisture storage;
- An increase in the flow variability of dryland rivers and streams;
- A decline in soil organic matter and nutrient status with an attendant decline in crop and fodder yields and, in extreme cases, social disruption, famine, and human and livestock migration;
- Salt accumulation in the surface horizons of dryland soils;
- An increase in the salt content of previously freshwater lakes, wetlands and rivers;
- The replacement of forest or woodland by secondary savanna grassland or scrub;
- A reduction in species diversity and plant biomass in dryland ecosystems.
- An increase in dust particles;
- An increase in carbon particles and trace gases from biomass burning.

24.7 Fire and the impact of biomass burning

Fire is often cited as a major cause of land degradation, and there is little doubt that during drought years the impact of wildfires can be devastating, as witnessed in recent years in California, northern Russia, China, the Mediterranean countries of Europe and most parts of Australia. The impact of unregulated fires throughout the drier parts of the world has now become so severe in terms of loss of life, livelihood and homes that it merits our particular attention.

We saw in [Chapter 17](#) that our prehistoric ancestors may have begun to use fire about a million years ago. Some of the evidence remains equivocal, but by Upper Palaeolithic/Late Stone Age times, fire was widely used for hunting, for preserving meat, for day to day cooking and for heating certain types of stone in order to make them easier to flake and fashion into knives and other implements. The Aboriginal inhabitants of Australia made frequent use of fire and contributed to substantial changes in the plant cover as well as the possible extinctions of certain species of

rainforest trees that were especially sensitive to fire, such as the *Auracaria*. To this day, the Aboriginal people of Arnhem Land in seasonally wet tropical northern Australia burn off the dead grass and undergrowth every year at the end of the wet season (usually late March or early April) in order to allow the unimpeded growth of new grass to attract wallabies and similar game (Haynes, 1991).

So widespread throughout Australia was the Aboriginal use of fire to modify the plant cover in ways they deemed desirable that the archaeologist Rhys Jones coined the expression *fire-stick farming* for this activity (Jones, 1968), although farming in the strict and narrow sense of cultivating domesticated plants is hardly an accurate term for this process. One of the great puzzles in the Australian prehistoric record is the demise of the megafauna roughly 45,000 years ago, at about the time that prehistoric humans first entered Australia (see Chapter 17). Although it is still unclear whether humans were the sole or main cause of these extinctions, it is possible that widespread prehistoric burning may have altered the plant cover and habitats of many of the large browsers and so hastened their demise. A corollary to this argument is the suggestion put forward by Miller and his co-workers that the change in vegetation led to a change in run-off and hydrological regime and even weakened the summer monsoon (Miller et al., 2005). In common with Charney's 'biogeophysical' drought model discussed in Chapter 23, this hypothesis is more of a numerical sensitivity study than a demonstration of an actual synoptic effect.

In the normal course of events, natural fires triggered by lightning strikes are most common at the start of the summer wet season and frequent low-intensity Aboriginal fires are most common at the close of the wet season. In either case, the fires are limited in their impact and seldom spread out of control. This changed with the arrival more than two hundred years ago of the Anglo-Celtic settlers, who rapidly excluded the first Australians from their expanding territory and in the process brought to an end the practice of frequent low-intensity firing of the landscape. As a consequence, the long-established fire regime was altered quite drastically, so frequent low-intensity burning became rare except in the far north of the continent, giving way to less frequent but high-intensity and highly destructive fires (Gammage, 2011). This pattern is also evident in other parts of the world, such as North America, where attempts to exclude fire and promote forest protection allowed a substantial build-up of the fuel load during wet years, leaving the forests increasingly vulnerable to lightning strikes or arson.

Fire can be a potent cause of land degradation and loss of biomass and biodiversity, even in the wetter parts of the dry subhumid regions of the world. Table 23.3 (Chapter 23) shows that during the 1877–1998 period, ENSO events coincided with both droughts and fires in Indonesia. About 12.5 per cent of Asia experiences biomass burning on an annual basis, although it is more severe and widespread during drought years. Some of the burning results from shifting cultivation, some from the clearing of forest for oil palms and other plantation products, and some from savanna fires

that were either deliberately lit or triggered by lightning strikes towards the end of the dry season. The total amount of carbon released from all sources of biomass burning up until 1990 was estimated as 500–1,000 Tg of carbon/year (Crutzen and Andreae, 1990), of which more than one-tenth came from Asia, one-quarter came from South America and two-fifths came from Africa (Andreae, 1993). (One tera-gram is 10^{12} g.) Fine particulate emissions from Asia amounted to $1.7\text{--}5.6 \times 10^{12}$ g/year, soot emissions equalled 0.4×10^{12} g/year and the total biomass burned 351×10^{12} g/year (Ghan and Penner, 1992). All of these figures have increased since then. Ghan and Penner (1992) estimated that total global fine particulate emissions from fossil fuel combustion amounted to $22.5\text{--}24.0 \times 10^{12}$ g/year, compared to $25.0\text{--}79.0 \times 10^{12}$ g/year from biomass burning. If droughts become more frequent even beyond the presently drier parts of Asia, as they are predicted to do (IPCC, 2007a), the health hazards from smoke inhalation will increase in the region centred on Indonesia-Papua New Guinea-Malaysia and Singapore, which shows that the effects of desertification processes can have repercussions well beyond the source area.

One important but often overlooked consequence of increased biomass burning in the seasonally wet tropics is the loss of the protective surface layer of vegetation, which reduces the impact of falling raindrops at the beginning of the wet season, leading to an order of magnitude increase in soil loss on even very gentle slopes (Chapter 10). The loss of soil will further reduce plant growth, leading to the onset of desertification. In fact, the French forester Aubréville (1949) first coined the term ‘desertification’ after many years of observing that biomass burning in tropical Africa was rapidly destroying the original forest cover and was creating desert-like conditions in areas with a mean annual rainfall of 750–1,500 mm/year. Needless to say, many savannas are not the result of the destruction of forest and woodland through the repeated and indiscriminate use of fire, since they reflect the primary impact of a seasonal rainfall regime and have been in existence for many millions of years before ancestral fire-using humans first appeared on the planet (see Chapter 3).

We now turn to examples of desertification in Australia, Africa and Asia in order to illustrate some of the general propositions outlined earlier. Australia is a fitting place to begin, because unlike every other inhabited continent, it evaded the impact of ten thousand years of Neolithic herding and farming until a mere two centuries ago, and has suffered from the resulting culture shock ever since.

24.8 Desertification in Australia

The Neolithic revolution (see Chapter 17) came late to Australia, but when it did come some 200 years ago, it had an impact that resonates to this day. Occupied for more than 40,000 years by small bands of hunter-gatherers, Australia had escaped the early impact of herds of sheep, goats and cattle, with their hard hooves and occasional destructive grazing habits. The first non-Aboriginal settlers, whether they

were soldiers or convicts, came from Britain and Ireland and were imbued with methods of farming adapted to those temperate and well-watered lands but poorly suited to the inherently infertile soils and highly variable rainfall regime characteristic of much of Australia. In their efforts to recreate the 'green and pleasant land' they had left, the first Anglo-Celtic farmers set about clearing the woodlands and forests with enormous energy and efficacy – a tradition that continues to this day, despite its destructive impact. In *A Land Half Won*, historian Geoffrey Blainey describes a scene at a farm in Gippsland, Victoria, about 150 years ago. The dying farmer summons his sons, points proudly to the total lack of trees where once stood forest and declares that he can now die content, his work well-done (Blainey, 1980).

It is easy to stand in judgement on the tree clearing activities of the early Anglo-Celtic settlers in Australia, but doing so has the wisdom of hindsight. The first European farmers probably had no idea of the adverse long-term effects of clearing the deep-rooted eucalyptus trees. The Polish-born geographer, naturalist and explorer, Paul de Strzelecki (1797–1873), who named Australia's highest mountain, Mount Kosciuszko, and after whom is named one of Australia's sand deserts, the Strzelecki Desert, was a perceptive early observer of accelerated soil erosion, warning against the adverse effects of clearing and burning on soil organic matter, soil structure, infiltration capacity and soil erosion (Strzelecki, 1845). However, even a man as observant and well-informed as Strzelecki did not realise that the eucalypts functioned as natural groundwater pumps and prevented the groundwater from rising as far as the rooting zone of shallow-rooted cereal crops such as wheat and barley, bringing dissolved salts to the surface in a process now known as *dryland salinization* (Macumber, 1991; Williams, 2000a; Lawrie and Williams, 2004). Under native vegetation, groundwater recharge in the Murray-Darling basin of south-east Australia amounts to a mere 1–2 mm/year, as opposed to 40–140 mm/year under wheat cultivation. Once the land was cleared, the groundwater levels rose, slowly but inexorably, bringing dissolved salts to the surface. There are 10^{11} tons of salt in the groundwater of the Murray Basin, and about 1.5 million tons of new salt are deposited into the Murray-Darling catchment each year from rainfall (Herczeg et al., 2001).

The loss of woodland and forest is not always easy to gauge, but best current estimates are that half of the original woodland and forest that grew in Australia 200 years ago are gone. Government policies aggravated the problem, because until quite recently, farmers were required by law to clear a certain area of their land each year or else forfeit tenure of the land. Sadly, the orgy of deforestation has not yet abated, and inconsistent and feebly enforced State policies offer little solace. In the ten years before 1993, Australia cleared an average of 500,000 ha of woodland and scrub each year, which some observers in this nation of sport enthusiasts have expressed as equivalent to two football pitches per minute. (The area of a standard football pitch is about 9,000 m² and that of an Australian Rules football playing field is about 20,250 m², so such comparisons are suggestive rather than precise, but

they do underline the speed of deforestation.) In 1990, Australia cleared a further 650,000 ha. According to Anderson (1995, p. 12), this amounted to ‘more than half the area cleared in the Amazon Basin’.

The resulting rise of saline groundwater has led to major loss of arable land across southern Australia. Western Australia had lost 1.6 million ha of farmland to salt by 1994, equivalent to about one-tenth of the cleared agricultural land in that state, and is losing an area the size of a football pitch every hour, with concomitant loss of many hundreds of species of plants unique to that state. In the Murray-Darling Basin (the agricultural heart of the nation), by 1992, more than 200,000 ha had been lost to salt. By 1996, South Australia had lost 400,000 ha and Victoria had lost 150,000 ha of arable land (Commonwealth of Australia, 1996). Lost agricultural production was estimated at \$500 million each year, but a later audit showed this to be a gross underestimate. Publication of the *Salinity audit of the Murray-Darling Basin* in October 1999 (Murray-Darling Basin Ministerial Committee, 1999), together with the CSIRO Land and Water Report (1999) on *Effectiveness of current farming systems in the control of dryland salinity*, revealed that more than 2.5 million ha of former agricultural land in Australia had become unusable because of dryland salinity. Cost to the Australian economy amounted to at least \$1 billion – more than double previous estimates. Unless current trends can somehow be curbed, more than 15 million ha will be salt affected over the next fifty years. More insidious is the damage to infrastructure from salt, with more than eighty towns in regional areas now threatened. The city of Adelaide in South Australia has more than a million people and depends primarily on the Murray River for its drinking water. In the Murray-Darling Basin, rising groundwater brings 5 million tonnes of dissolved salt to the surface each year, with 2 million flowing down the river and 3 million remaining on the land preventing farming.

Dryland salinization is not the only problem of land degradation facing Australia; nor are all the areas of saline land a result of human action – many parts of the continent have been accumulating salt throughout geologically recent times (see Chapter 22). Lake Eyre is a case in point. Some of the salt problems are caused by irrigation, although in general this is well-managed. However, excessive abstraction of water upstream for growing cotton or rice can seriously deplete the supplies of freshwater available downstream and can lead to build-up of salt in the distal parts of drainage basins like the Murray. Much of this salt is cyclic, that is, blown in from the surrounding oceans, but some comes from times when the sea invaded the land at intervals during the Neogene.

Another issue is that of localised overgrazing of rangelands in much of central Australia, coupled with the effects of a change in fire regime on the native plant cover. Some ingenious methods have been devised in an attempt to separate out the effects of drought from those primarily caused by overgrazing (Pickup, 1996; Pickup, 1998), but these require detailed ground control calibration to be effective, which is expensive

Table 24.2. *Extent and severity of desertification in irrigated areas, rain-fed croplands and rangelands in the areas classed as drylands in Asia (in Thousands of Hectares). (From UNEP, 1992b.)*

Areas shown	Total area	Slight to none	Moderate	Severe	Very severe	Total Moderate or worse	% degraded
Irrigated area	92,021	60,208	24,335	5,788	1,690	31,813	35
Rain-fed cropland	218,174	95,890	100,638	18,578	3,068	122,284	56
Rangeland	1,571,240	383,630	485,221	691,602	10,787	1,187,610	76

and so far quite rare. A related problem is the lack of a unified national strategy for monitoring and implementing optimal land management procedures, given that responsibility still lies in the hands of individual state agencies, not all of which are able to perform these roles wisely and efficiently.

24.9 Desertification in Asia

The most comprehensive maps of the present extent of global and regional land degradation and desertification so far available are those published by UNEP (1992a; 1997). In Asia as a whole, 56 per cent out of a total area of 218.2 million hectares of rain-fed cropland were classed as degraded (Table 24.2), of which 122.3 million hectares were ranked as being moderately to very severely affected (UNEP, 1992b). Corresponding figures for rangelands in Asia (1,571.2 million hectares in total) amounted to 76 per cent, or 1,187.6 million hectares, being moderately to very severely affected (UNEP 1992b). Irrigated areas in Asia covered 92 million hectares, with 35 per cent, or 31.8 million hectares, being moderately to severely affected (UNEP, 1992b). In all cases, the arid and semi-arid areas were the most severely affected, but the dry sub-humid areas were most vulnerable to future change, since they were far more densely populated and had less collective experience in adapting to extreme climatic vicissitudes. Within the arid and semi-arid regions, there was often a considerable body of detailed information available to assist pastoralists in managing their resources more effectively, as in the Cholistan Desert of Pakistan (Mohammad, 1989; Rao et al., 1989).

Salinization has been a major problem in the Indus Valley since before the 1950s and has become more evident in the drier parts of China in the last twenty years. In both cases, the root causes were over-irrigation, inadequate drainage and locally rising groundwater-tables. One concern is what the impact of current warming trends in this region might be. Higher temperatures will probably lead to an increase in water losses from crops through transpiration and from canals through evaporation, leading to

further demands on already overstretched irrigation systems. A related issue is water quality, which requires basin-wide biological and chemical monitoring programmes to be implemented, as in the exemplary scientific study of the Ganga River in northern India (Krishna Murti et al., 1991).

Menon et al. (2002) have modelled the present-day consequences of the increase in atmospheric carbon particles over China (and, possibly, India), concluding that they will lead to an increase in floods in the south of China and an increase in drought and desertification in the north (Menon et al., 2002; IPCC, 2007a). However, this must remain conjectural until all other factors are considered.

The following two regional examples of desertification in Asia are given in some detail because they illustrate how human activities can aggravate the effects of an already dry climate, with the rider that increasing evaporation and less reliable rainfall or river flow postulated for the future (see Chapter 25) will exacerbate desertification processes.

24.9.1 Central Asia: drying up of the Aral Sea

The drying up of the Aral Sea has been variously described as ‘the world’s largest man-made disaster’ (Tanton and Heaven, 1999) and ‘perhaps the most notorious ecological catastrophe of human making’ (Stone, 1999). The volume edited by Glantz (1999) avoids such hyperbole and provides a comprehensive overview of previously relatively inaccessible hydrological, ecological and human health investigations within the Aral Sea Basin, and it complements earlier studies of environmental degradation within the basin (Babaev, 1996; Micklin and Williams, 1996).

Two major rivers flow into the Aral Sea from the east: the Amu Darya and the Syr Darya. The present Aral Sea is the remnant of a once extensive early Holocene lake (Glazirin and Trofimov, 1999; Boomer et al., 2000). Very gradually, over the previous half-century, the environmental problems in the basin worsened. Glantz (1999) terms this phenomenon ‘creeping environmental change’ and defines it as the result of ‘long-term, low-grade, incremental but cumulative environmental problems’. The origin of the crisis began early last century with the search for self-sufficiency in cotton from Central Asia, and it was aggravated by the construction of the 1,400-km long Karakum Canal, which deprived the Aral Sea of 15 km³ of water each year. The outcome was stark, with what was once the fourth largest lake in the world shrunk to less than half its former size and its volume reduced by more than two-thirds.

The consequences for the surrounding region (*Priaraliye* in Russian) have been analysed in a number of perceptive studies (Saiko, 1998; Spoor, 1998; Glantz, 1999; Saiko and Zonn, 2000). They include exposure of saline ground, contamination of water and soils with salts and pesticides, accelerated soil loss in dust storms, collapse of the regional fishing industry, and a rapid and widespread decline in human health.

Recognition of the problem has brought belated remedial measures, including repairs to the banks of the Syr Darya to reduce seepage losses and the construction of a dyke to enable more water to remain in the northern sector of the Aral Sea (Pala, 2008). The success of these measures will depend on the more economical use of water and especially on whether climate change in the headwaters of the two main rivers will lead to reduced flow and increased evaporation. Similar caveats apply to the recent plans by Turkmenistan to fill the Karashkor Depression with surplus drainage water, which may contain high levels of dissolved pesticides and herbicides (Stone, 2008). Any increase in climatic variability (as foreshadowed by the IPCC, 2007a) will accelerate human migration out of these parts of Central Asia. Creeping environmental change in the Aral Sea Basin over the past fifty years may offer a useful analogue for the present and future impact of climatic change in the desert world.

24.9.2 China: dune reactivation in northern China

The Desert Research Institute of the Chinese Academy of Science in Lanzhou also has field stations at Shapotou and Yanchi and has over the years published excellent data on the extent and severity of desertification in China (Zhu et al., 1989; Zhu and Wang, 1992; Zhu et al., 1992; Wang, 1993a; Wang, 1993b; Zhu and Wang, 1993). The *Atlas of Natural Disasters in China* (Zhang et al., 1992) provides useful additional information. The Chinese government has long been acutely aware of the need for desertification control (Ci, 1998), and the increase over the past two decades in dust storms afflicting Beijing has reinforced this perception. China's Agenda 21 (1994) devotes a full chapter to desertification, and the *Beijing Review* publishes thoughtful articles on the causes and economic impacts of desertification in China, but the possible impacts of future climate change are perhaps insufficiently embedded in current strategic planning, despite Chinese scholars having long been aware of the impact of historic changes in climate on desertification in Xinjiang and further to the east (Zhang et al., 1991; Xia et al., 1993; Xia et al., 1995; Zhang, 1995).

China is unique in having the largest and thickest loess mantle of anywhere in the world, and the highly fertile loess soils are peculiarly susceptible to accelerated erosion, resulting in enormous volumes of sediment carried in suspension by the big rivers draining the Loess Plateau. (As an aside, more than two-thirds of the 20 billion (10^9) tonnes of sediment carried each year to the world's oceans comes from five river basins, two in China – the Yangtze and Huanghe – and the other three from the Himalayas via India, Bangladesh and Pakistan, respectively – the Ganga, Brahmaputra and Indus.)

The loess in China was laid down during colder, drier and windier climatic intervals in the geologically recent past, and it became vegetated and weathered to form soils during warmer, wetter phases when the summer monsoon was strong (Chapter 9). This

loess is now becoming eroded through excess clearing and overcultivation, a familiar practice in the history of land degradation that is worse in times of drought and was responsible for dynastic change in the past. Despite some substantial achievements in desertification control, rates of soil and nutrient loss from the Loess Plateau remain very high (Douglas, 1989; Liu, 1999; Fu et al., 2000).

Equally important is the clearing of once vegetated and stable dunes in northern China; what is known as sandy desertification in that region increased very rapidly from 176,000 km² in the mid-1970s to 197,000 km² in the mid-1980s, a mean yearly increase of 2,100 km² (Wang, 1993b). Of the major human causes of sandy desertification in northern China in the early 1990s, Wang (1993a) estimated that in terms of total area affected, over-collection of wood for fuel accounted for 31.8 per cent, overgrazing on the steppes for 28.3 per cent, overcultivation on the steppes for 25.4 per cent and misuse of water resources 8.3 per cent, with other factors responsible for the balance of 6.2 per cent. However, there is a clear need to re-evaluate earlier estimates of the extent and severity of desertification in northern China (Williams, 2000a).

One region of particular concern at present is the Inner Mongolian Autonomous Region, especially the Alashan Plateau in the north-west of the region, where there is an accelerating influx of wind-blown sand from once stable dunes into the Huanghe (Yellow) River. The Alashan region of Inner Mongolia is one of the driest and poorest regions in China. It covers an area of roughly 270,000 km², with rainfall declining from 300 mm in the east of the region to less than 50 mm in the west (Yang, 1991). Mountains occupy about 10,000 km² and are flanked by gently sloping sand and gravel alluvial plains, or *gobi*, which cover 91,000 km² of the Alashan. In the north-east, west and south-east of the Alashan, there are three major active dune fields that occupy an area of about 81,000 km². Fixed and semi-active dune fields cover roughly 90,000 km² and are the places most vulnerable to desertification.

Until the 1950s, many of these low dune fields and sand sheets were covered in a relatively dense cover of shrubs, trees and grasses. Since that time, the human population has doubled and livestock numbers have tripled, mainly as a result of an influx of immigrants from Gansu Province and the Ningxia Autonomous Region to the south (Williams, 2000a). In addition, there have been a number of severe droughts, including the exceptionally severe 1989 drought (during an El Niño year). Local Mongolian farmers considered the unusually heavy rains of 5 and 6 August 1999 (a La Niña year) to be the best rains of the previous decade, bringing the preceding three-year drought to an end (Williams, 2000a).

The combination of sporadic but severe droughts much increased stock numbers, and the influx of economic refugees from the south has led to widespread and locally severe overgrazing, the destruction of localised patches of forest and severe desertification. Local estimates in July 1999 suggested that 30,000 km² were severely degraded, with the rate of desertification increasing by about 1,000 km² each year



Figure 24.3. Abandoned homesteads on the Alashan Plateau, Inner Mongolia, northern China. Accelerated erosion followed a sudden increase in the human and animal population in the 1950s as a result of migration from the south.

(Williams, 2000a). These are probably reasonable order of magnitude estimates, given the lack of accurate soil and vegetation maps against which to assess them.

Discussion with farmers and herdsman before the drought broke in August 1999 revealed widespread concern over a perceived increase in sand and dust storms and a decline in pasture quality (Williams, 2000a). There was abundant and unambiguous evidence of accelerated soil erosion by wind and water since the 1950s. Former agricultural settlements immediately west of the rugged and wooded Helan Shan range have been abandoned (Figure 24.3), and the adjacent fine-grained alluvial soils have been severely gullied. In some places, only a single tree remains as a witness to the riparian woodland of the 1950s.

Of even greater concern was the reactivation of previously stable and vegetated dunes to the north and west of the mountains as a result of overgrazing by goats and sheep. Monitored rates of dune advance amounted to more than 10 m/year near the Yellow River to less than 1 m/year further inland. Along an 80 km reach on the left bank of the Yellow River, opposite the industrial city of Wuhai, the dunes were advancing at rates of up to 10 m/year. An estimated 80 million m³ of sand is being blown into the river each year in this sector. Major reforestation of the mobile

dunes is underway, and a useful start has been made at the Shu Gui Desert Control Station, although tree losses from insect pests and groundwater salinity had been substantial. Control of desertification along the Yellow River is an important element of China's national plans to combat desertification (China's Agenda 21, 1994, p. 189). Overgrazing and dune reactivation are the prime causes of the present massive influx of sand into the Yellow River, but unless upstream abstraction of water for irrigation is curtailed, the silting up of the river will inevitably become worse.

If rainfall in northern China becomes more erratic than it is already (Ye et al., 1987; Domrös and Peng, 1988), as the IPCC forecasts suggest (IPCC, 2007a; IPCC, 2007b; IPCC, 2007c), there is likely to be further movement of people and livestock into areas perceived as underpopulated, with concomitant destruction of the plant cover and reactivation of the still vast areas of fixed dunes. In northern Xinjiang, many of the farmers harvest meltwater from the snow-capped Tian Shan ranges for irrigating their fruit and vegetable plots. As warming proceeds, the ablation of snow in summer will exceed winter accumulation, glaciers will continue to retreat and the reliable supplies of irrigation water from springtime snow-melt will diminish.

Any reduction in plant cover in semi-arid or seasonally wet tropical areas leads to a dramatic increase in sediment loss, with concomitant loss of arable land, pasture and woodland resources and often massive sedimentation problems downstream, as in the case of the Loess Plateau of China and the Huanghe, or Yellow River. Human migration from areas of high to low population density in parts of northern China since the 1950s has had a discernible and severe impact on the soils and plant cover, which has led to an increase in sand and dust storms. It is as yet hard to foresee whether these conditions will become more widespread throughout the region.

24.10 Desertification in Africa: an example from Ethiopia

The recurrent twentieth-century droughts in Africa caused widespread social distress, famine and migration, and proved to be the catalyst for the Desertification Convention discussed in Section 24.11. The droughts and famines in Ethiopia in particular have attracted considerable media attention, and the widespread political unrest generated by recurrent famines in the 1970s and early 1980s helped bring to an end the long reign of the Emperor Haile Selassie and that of the military dictator Haile Mariam Mengistu. Ethiopia is well-endowed with fertile volcanic soils and with large perennial rivers, so at first blush it seems strange that famines are so frequent. One reason that is often advanced for the low crop yields obtained by many farmers is indiscriminate deforestation and accelerated soil erosion. How valid is this perception?

It certainly seems true that in the Ethiopian Semien Mountains National Park, the area of natural forest declined from 56 per cent to 22 per cent within forty years (Hurni, 1999). Clive Nicol, the first game warden appointed by the Imperial Ethiopian Government to manage the park, found that he faced an impossible task in seeking

to prevent deforestation in the earliest years of the park's existence. He wrote despairingly of the resultant destruction of the once prime forest:

Seen close, the destruction was incredible. The place looked like a First World War battleground. Everywhere the ground was strewn with hot ash, smoking debris, charred stumps, and partially burned tree trunks, lying about willy-nilly. Some of the felled trees had been sixty feet [about 20 m] and more in height, and now they lay burned on the ground. I could hardly believe that a few men with simple, blunt iron axes had felled so many huge trees. Cedar, olive, hagenia, Podocarpus, euphorbia and many others. This year alone, Bogale and his sons and followers had cut and burned about three square kilometres of forest. (Nicol, 1971, p. 262)

Removal of forest can alter the local hydrological balance, leading to increased run-off, reduced infiltration and the drying up of springs and stream headwaters once fed by base flow and by the subsurface lateral flow of the infiltrating soil water, as discussed in [Chapter 10](#). A related effect is accelerated movement of soil downslope by slopewash, soil creep and landslides. The Swiss geographer Hans Hurni investigated rates of soil erosion in the Ethiopian Highlands (Hurni, 1999). He found that mean rates of soil loss amounted to 40 t/ha/year on uncultivated mountain slopes and rose to more than 300 t/ha/year when cultivated, or five to ten times more than those in cultivated areas beyond the mountains. He also observed a reduction in the length of fallow to virtually zero and an expansion in the area under cultivation. In part of Gojjam Province, the area cultivated rose from 40 per cent in 1957 to 77 per cent in 1995, while natural forest decreased from 27 per cent to 0.3 per cent (Hurni, 1999).

It is instructive to compare current rates of soil loss in the Ethiopian headwaters of the Blue Nile and Atbara (known in Ethiopia as the Abbai and Tekazze rivers, respectively) with long-term geological rates of denudation, expressed as a mean rate of surface lowering. The mean rate of surface denudation over the past 20 million years amounts to 0.01 mm/year (McDougall et al., 1974; Williams and Williams, 1980), which is an order of magnitude lower than the mean rate of 0.12–0.24 mm/year in the seventy years before completion of the Aswan High Dam in 1970. In the 1970s, soil loss from parts of the Ethiopian Plateau amounted to 0.4–1.0 mm/year (Nyssen et al., 2004).

The effects of such accelerated soil erosion have repercussions well beyond the local area. For example, before completion of the Sudanese Roseires Dam on the Blue Nile close to the Ethiopian border in 1965, Blue Nile suspended sediment load amounted to $50\text{--}100 \times 10^6$ t/year when measured at Khartoum but had fallen to $40\text{--}70 \times 10^6$ t/year from 1967 to 1969, largely as a result of silt trapped upstream of the dam (El Badri, 1972). By 1996, the capacity of the Roseires Reservoir had been reduced by nearly 60 per cent and that of the Khashm el Girba Reservoir on the Atbara by 40 per cent (Swain, 1997). (The Atbara rises close to the Blue Nile headwaters in Ethiopia, where it is known as the Tekazze). It appears that deforestation in and around the Semien Mountains during the past half-century has initiated a pulse of accelerated

soil erosion from cultivated land and has led to silt accumulation in reservoirs many hundreds of kilometres downstream.

It is easy to exaggerate the effects of deforestation by extrapolating from a few well-studied areas to an entire region or country. A number of careful studies of forest clearing and regrowth in other parts of Ethiopia have shown that in some areas, forests have now replaced land that had been cleared and cultivated as recently as half a century ago (McCann, 1999, pp. 93–104). Indeed, as the author has noted elsewhere: ‘The concept of recent, progressive and linear deforestation in the Ethiopian highlands is therefore open to question’ (Williams, 2000a, p. 241). Likewise, in regard to land degradation in Ethiopia, Kenya and Zimbabwe, detailed agronomic studies have shown that soil fertility decline may indeed be true of some farmers at certain places and times, but it is far from universal (Leach and Mearns, 1996; Scoones, 1997; Elias and Scoones, 1999; Bassett and Crummey, 2003).

24.11 Global estimates of the extent and severity of desertification

There are several reasons why it is very hard to estimate the extent and severity of desertification on a regional and global scale. One reason is that desertification involves a complex variety of processes, not all of which are immediately obvious. Another relates back to the protracted and continuing debate over the relative importance of human and natural factors in causing land degradation. Yet another reason is the simple lack of data and effective monitoring protocols, especially but not solely in many of the poorer countries of the world. Finally, it has been difficult to reach agreement about the most appropriate criteria with which to define the extent and severity of desertification processes. It is unlikely that we will achieve uniform methods and diagnostic criteria to evaluate desertification processes in the immediate future, and in the absence of universally accepted criteria, it will be no easy task to monitor current trends.

Despite this unpromising context, there have been a number of attempts to assess global land degradation in drylands. UNEP attempted this exercise in 1977, 1984 and 1987 (UNEP, 1992b). Some of these estimates came from questionnaires sent out to government agencies already confronted with drought, famine and widespread social distress, and they were probably neither precise nor accurate (Binns, 1990; Rhodes, 1991; Thomas, 1993). Dregne et al. (1991) revised these estimates, concluding that three-quarters of all rangelands on every continent were to some extent degraded, together with 15–30 per cent of irrigated lands, half of the rain-fed croplands in Africa, Asia and Europe, more than one-third of those in Australia and South America and less than one-fifth of those in North America. However, even these figures should be treated as guesstimates (Harold Dregne, personal communication, 1993).

In an effort to obtain more credible and more quantitative estimates of land degradation, UNEP entered into a partnership with the International Soil Reference Centre

Table 24.3. *Extent of soil degradation in susceptible drylands (in Millions of Hectares) (From UNEP, 1997)*

Region	Type of soil degradation				Total
	Water erosion	Wind erosion	Chemical deterioration	Physical deterioration	
Africa	119.1	159.9	26.5	13.9	319.4
Asia	157.5	153.2	50.2	9.6	370.5
Australia	69.6	16.0	0.6	1.2	87.4
Europe	48.1	38.6	4.1	8.6	99.4
North America	38.4	37.8	2.2	1.0	79.4
South America	34.7	26.9	17.0	0.4	79.0
Total	467.4	432.4	100.7	34.7	1035.2

(ISRIC) in the Netherlands, and in 1987 both partners began the Global Assessment of Soil Degradation (GLASOD) project. Data from this project were used in both editions of the *World Atlas of Desertification* (UNEP, 1992a; UNEP, 1997). Table 24.2 shows the global extent and severity of soil degradation obtained from the GLASOD data (UNEP, 1997). Nevertheless, we should only regard these estimates as order of magnitude values, since in most instances we still lack adequate ground control. For example, it seems highly likely that the extent of salt-affected soils in Australia is seriously underestimated in Table 24.3. There is also a methodological problem in defining areas by drawing polygons around what are often only point sources of soil degradation, leading to overestimates (David Thomas, personal communication, 2000). This is particularly the case in regard to overgrazing around wells, boreholes and other watering points in the rangelands of Africa and Australia (Pickup, 1996; Pickup, 1998). Despite these methodological teething problems, it seems reasonable to accept that the degradation of dryland soils and ecosystems is a major problem on every inhabited continent in terms of both the total areas affected and the relative proportions of arable and pastoral lands.

24.12 Mitigation and prevention of desertification

Desertification is caused, directly and indirectly, by a variety of biophysical, social and economic factors and therefore needs to be considered at a variety of scales ranging from local through regional and national to international. We start with the international efforts to curb or prevent desertification. One potentially important initiative to emerge from the 1992 Earth Summit was a request to the UN General Assembly to establish an intergovernmental committee for negotiating a convention on desertification. At its forty-seventh session in 1992, the General Assembly acted swiftly and set up the International Negotiating Committee, which met five times to debate

the complex issues involved. The negotiations prospered, and eighty-seven nations signed the Desertification Convention in Paris on 14–15 October 1994, a remarkable achievement given the diversity of opinions involved (Kassas, 1995b; Williams et al., 1995; Chasek, 1997). The full and rather cumbersome title of this convention is the *United Nations Convention to Combat Desertification in those countries experiencing serious drought and/or desertification, particularly in Africa*. We have discussed the reasons for the emphasis on Africa and for the mention of drought in the Introduction to this chapter. A major goal of the Desertification Convention was to encourage individual countries to prepare their own action plans to combat desertification. (Use of the bellicose verb ‘to combat’ seems, in the author’s view, to give the wrong emphasis. Prevention and management would be more appropriate terms).

More specific discussion of some practical measures to alleviate desertification and of some key principles upon which to achieve sustainable use of desert environments is best deferred until Chapter 26, after we have considered current climatic trends and possible future changes. However, some general comments can still be made.

The spread of apparently unpalatable and ‘useless’ weeds into areas of previously good grazing is not always an ecological disaster. Desert plant ecologists have long been concerned with the impact of grazing on species diversity. In an elegant study of the role of the spiny cactus *Opuntia polycantha* as a plant refuge in grazed, short-grass, steppe-plant communities in North America, Rebollo et al. (2002) found greater species diversity among the clumps of cactus than away from those plants, which means that a spiny cactus that some might consider a weed in fact plays a useful role in maintaining species diversity.

The success or otherwise of the many measures designed to reduce or prevent desertification processes is seldom publicised, so it is often hard to determine the degree of success of prevention and rehabilitation strategies. Any cost-benefit evaluation of successful projects should take into account the fundamental principles of long-term, ecologically sustainable development rather than focussing solely on short-term and purely economic criteria.

Given the lack of primary meteorological, agrometeorological and hydrological data for many of the drier regions of Africa, Asia and South America, it would be useful to establish regional climate monitoring stations in those areas where the human population is likely to be most vulnerable to probable future changes in climatic variability. Recognising the need for the enhanced provision of information governing more sustainable forms of land use, greater efforts should be made to provide information in an accessible and useable form to local farming and pastoral communities (Williams and Balling, 1996; Barakat and Hegazy, 1997; Imeson, 2012). The regional centres just mentioned could be appropriate means of providing such information to local communities. There are existing models from North America and Australia that could be adapted to perform this role.

24.13 Conclusion

A persistent problem in assessing the causes of desertification arises from the fact that it is often very hard to separate out the effects of climatic variability from those of human activities. However, the three examples discussed in some detail for Australia, China and Ethiopia indicate very clearly that overenthusiastic removal of the vegetation cover can lead to severe and widespread forms of land degradation, ranging from rising saline groundwater (Australia) to dune reactivation (northern China) to accelerated soil erosion (Ethiopian Highlands). In addition, the effects of such deforestation/removal of the plant cover can extend well outside the original area, leading to choking of rivers with sediment and siltation of reservoirs hundreds of kilometres downstream.

Analysis of the causes of land degradation processes reveals a complex web of interactions in which social and economic factors are often at least as important as purely biophysical factors. Looking to the future, those areas now most vulnerable to interannual climatic variability will continue to be the arid and semi-arid areas, but the seasonally wet tropics and the dry subhumid regions that now receive 1,500–750 mm of rain a year will become increasingly sensitive to any future changes in temperature, evaporation and precipitation, as discussed in [Chapter 25](#). Analysis of the causes and consequences of desertification processes related to both climatic variations and human activities suggests that human migration within and into these regions can seriously aggravate existing problems of land degradation.

Biomass burning in the seasonally wet tropics, notably the dry subhumid regions already severe during dry years, is generating health problems, aggravating soil erosion and has the potential to change regional climate. In addition, in the absence of universally accepted criteria to define the extent and severity of desertification processes, it is difficult to monitor current trends within the drylands of the world. Finally, the success or otherwise of the many measures designed to reduce or prevent desertification is seldom publicised, so that it is often hard to determine the degree of success of prevention and rehabilitation strategies.

Current climatic trends in deserts and possible future changes

Time present and time past
Are both perhaps present in time future,
And time future in time past . . .
T.S. Eliot (1888–1965)
Four Quartets: 'Burnt Norton' (1944)

25.1 Introduction

Climate change is nothing new, and humans have had a long history of adapting to the vicissitudes of such change (Braudel, 1949; Lamb, 1972; Lamb, 1977; Le Roy Ladurie, 1971; Dunnette and O'Brien, 1992; Mintzer, 1992; Whyte, 1995; Zhang, 1995). What is new, however, is the impact on global climate of the accelerating inputs of the greenhouse-enhancing gases, notably carbon dioxide, methane and nitrous oxide, all of which are causing warming of the lower atmosphere, the land and the sea (Berger, 1992; Graedel and Crutzen, 1995; Duplessy, 1996; Pittock, 2003; IPCC, 2007a; Houghton, 2009; Pittock, 2009; Cleugh et al., 2011; Dessler, 2012; Cowie, 2013; Eggleton, 2013).

Predicted impacts of such climatic change include an increase in extreme climatic events (Watson et al., 1998; Cubaschi et al., 2001; IPCC, 2007a; Houghton, 2009; Whetton, 2011; Dessler, 2012; Eggleton, 2013), possible changes in oceanic thermohaline circulation (Broecker, 1992; Broecker, 2000), aggravated water shortages in the drier parts of the world and in regions dependent on snow-melt (Beniston, 2002; IPCC, 2007b), and an increase in the magnitude and frequency of floods and droughts (Cubaschi et al., 2001; IPCC, 2007a; IPCC, 2007b) and of tropical cyclones (Warrick and Ahmad, 1996; IPCC, 2007a). All of these impacts will accentuate existing and long-term changes in global land use (Turner et al., 1990; Meyer and Turner, 1994) and will have increasingly severe repercussions for tropical forests (Sayer and Whitmore, 1991) and for global biodiversity (Heywood, 1995). The monumental survey edited

by Turner et al. (1990) covering three centuries of human impact on the earth contains many thoughtful and perceptive contributions to this important topic, embracing both the social and natural sciences. No comparable study has appeared since.

There is already concern that the magnitude and frequency of extreme climatic events, such as floods, droughts and heat waves, will increase during this century as a result of global climatic changes caused in large part by human activities (Cubaschi et al., 2001; Meehl and Tebaldi, 2004; IPCC, 2007a; IPCC, 2007b; Houghton, 2009; Dessler, 2012; Eggleton, 2013). Indeed, if present trends continue, some workers believe that those regions of the world already experiencing large variations in precipitation from year to year will be likely to experience an even higher degree of variability (Cubaschi et al., 2001), which will require adaptation and more flexible strategies of risk management among pastoral and agricultural communities. Williams and Nottage (2006) noted that the areas likely to be most affected are the arid, semi-arid and dry subhumid regions of the world that cover nearly 40 per cent of the total land area and comprise roughly one-third of the total world population. Most at risk will be the poorer communities in regions already vulnerable to land degradation and desertification (Williams, 2002b; Williams and Nottage, 2006).

Within this context, the aim of this chapter is to consider current trends in global and regional climate, especially precipitation and temperature, to assess some of the possible causes of these trends and to present some possible scenarios of future climate change and their likely impacts.

25.2 Causes of climatic change

‘Let us admit at once that we do not know what are the basic causes of climatic change’ (Flint, 1971, p. 789). More than forty years have elapsed since Flint, having reviewed at length the legacy of Quaternary glaciations on every continent and the evidence of past environmental changes in areas that had escaped glaciation, finally pronounced that the fundamental causes of climatic change still remained elusive. There has been a great deal of work on this topic in the years following the publication of Flint’s magisterial volume, including the analysis of sediment cores collected from the floor of the ocean (Shackleton, 1977; Shackleton, 1987; Shackleton, 2001) and of ice cores drilled from the Greenland ice caps (Dansgaard et al., 1984; Dansgaard et al., 1985; Dansgaard et al., 1993) and the Antarctic ice caps (Jouzel et al., 1995; Jouzel et al., 1997; EPICA, 2004; EPICA, 2006; Jouzel et al., 2007; Lüthi et al., 2008) and a number of smaller ice caps (Thompson et al., 1995; Thompson et al., 1998; Thompson et al., 2002).

In addition, technical advances in methods of dating climate proxies, such as lake, river and wind-blown sediments, as well as the development of high-precision chronologies from marine and glacial archives (see Chapter 6) have led to major advances in our reconstruction of the pattern and tempo of past climatic events. We now know a great deal more about a number of the factors that control climate change at different

Net global albedo = 30%

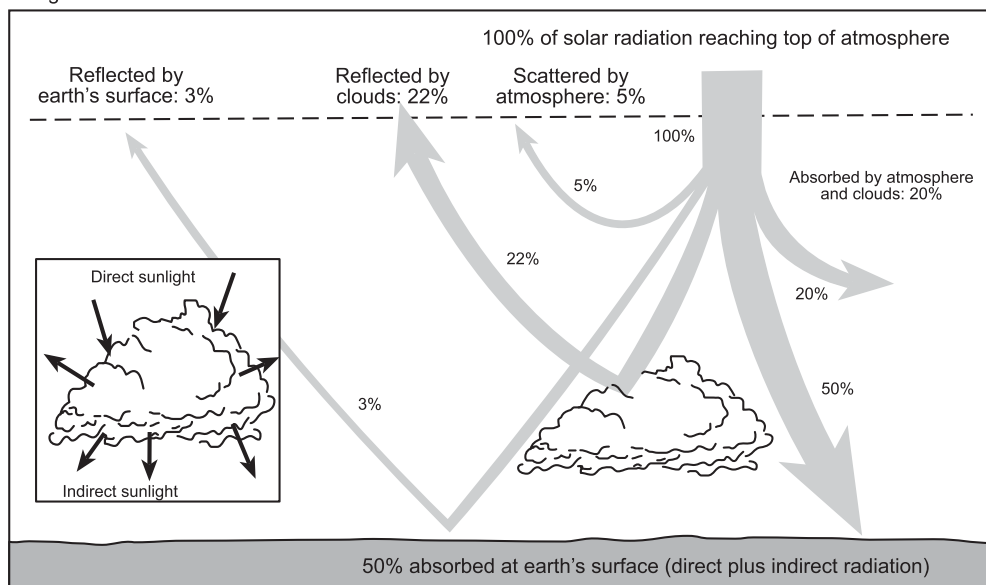


Figure 25.1. Solar radiation budget showing inputs and outputs. (Adapted from Anon., 1989.)

scales in time and space, so that Flint's overly pessimistic conclusion no longer seems valid. There is, of course, still scope for disagreement and differing interpretations of the primary data. (For a thoughtful analysis of the causes of such disagreement, see Hulme, 2009). Nevertheless, the quantity and quality of precise and accurate information available to us does allow some definite conclusions to be drawn about what we do know, what we do not as yet know and what we will need to know if we are to adapt intelligently to possible future change. Before we can consider the causes of climate change, we need to understand what factors govern the present-day climate.

25.2.1 Factors that determine present-day climate

Solar energy is the ultimate agent that determines our climate on earth. Solar energy is transferred to the earth by what may best be described as oscillating electromagnetic forces (radiation), which have different wavelengths (Anon., 1989). Most of the solar radiation that reaches the surface of the earth is in the form of visible light, with a wavelength of 0.4–0.7 μm . Solar energy penetrates the earth's atmosphere and reaches the earth's surface in the form of relatively short-wave solar radiation, which is then radiated back from the earth's surface as relatively long-wave terrestrial radiation. Only about 50 per cent of all solar radiation is received at the earth's surface (Figure 25.1). Roughly 20 per cent is absorbed by the atmosphere and by clouds, and

5 per cent is scattered by the atmosphere. A further 22 per cent is reflected back to space from the clouds, and 3 per cent is reflected back from the ground. (The global albedo, or reflectivity, therefore amounts to about 30 per cent).

The shortest path for solar radiation to reach the earth's surface is when it is aligned at right angles to the earth, which is at the equator. With distance away from the equator, solar radiation has to penetrate an increasingly thick layer of atmosphere, because the sun's rays are now aligned obliquely to the surface of the earth. This results in minimum warming in polar latitudes and maximum warming in tropical latitudes. The outcome of this differential heating, combined with the earth's rotation, results in two primary circulation patterns: the Hadley Cell and the Polar Cell ([Chapter 2, Figure 2.1](#)). Latitude is thus the primary factor responsible for global temperature distribution and the location of major frontal zones. Any changes in the latitudinal position of land-masses will have an impact on long-term climatic changes, as we saw in the case of the Cenozoic displacements of Africa and Australia ([Chapters 18 and 22](#)).

The seasonal displacement of these atmospheric circulation cells and frontal zones ([Chapter 2, Figure 2.2](#)) is a direct consequence of the tilt of the earth's axis, which causes an apparent northward displacement of the sun during the boreal summer and a southward displacement during the austral summer. Here again, long-term changes in the tilt of the earth's axis will cause long-term changes in seasonal climate (see [Chapter 3](#)).

Altitude is another important factor controlling temperature. Temperature decreases with increasing elevation at a rate of about $6^{\circ}\text{C}/1,000\text{ m}$, which is known as the lapse rate. The temperature difference between the lowest and highest points on earth amounts to about 55°C , which is about the same as the January mean sea level potential temperature variation with latitude (from $+20^{\circ}\text{C}$ to -40°C , or about 60°C). Over long time scales, tectonic uplift and subsidence will have an impact on local temperatures.

Finally, ocean circulation and distance inland ('continentality') determine the present-day temperature range on land, as well as moisture availability. Once again, any long-term changes in continental configuration will be apparent as long-term changes in precipitation and temperature. However, our concern in this chapter is not with the long-term changes, since we have already considered these in detail for all of the major deserts and their margins in [Chapters 18–22](#). Our concern here is with the more recent changes of the past few centuries.

25.3 The 'greenhouse effect'

We saw in the previous section that only about half of all solar radiation is actually absorbed at the earth's surface; the rest is either absorbed or scattered by the atmosphere and clouds, or else it is reflected back to space ([Figure 25.1](#)). The earth's atmosphere absorbs about 20 per cent of the outgoing long-wave (infrared) terrestrial

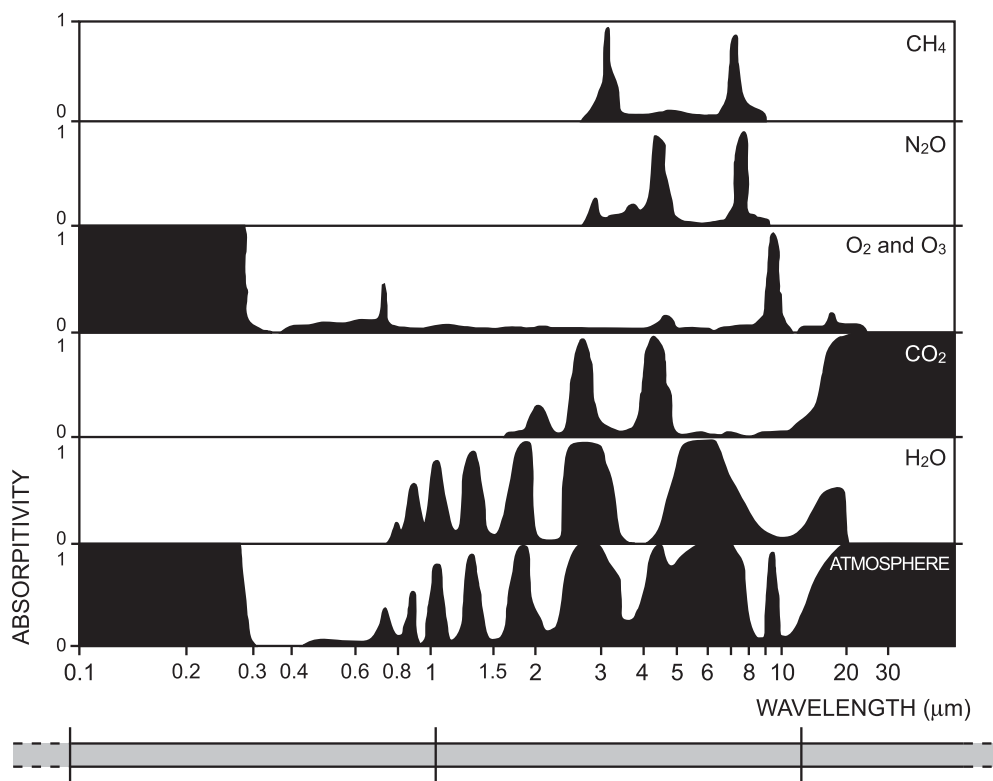


Figure 25.2. Absorption of solar radiation by certain atmospheric gases. (Adapted from Anon., 1989.)

radiation, and it therefore acts as a form of thermal blanket, keeping the earth warmer than it would otherwise be. If the earth did not have an atmosphere, the mean temperature at the surface of the earth would be -18°C , rather than the present $+14^{\circ}\text{C}$. The ‘greenhouse effect’ is a metaphor for this phenomenon. Our present atmosphere consists of 79 per cent nitrogen, 20 per cent oxygen and 1 per cent other gases. It is certain of these atmospheric gases that are so important in regulating earth temperature. They include water vapour (H_2O), carbon dioxide (CO_2), oxygen (O_2), ozone (O_3), nitrous oxide (N_2O) and methane (CH_4), as well as the entirely synthetic greenhouse gases: the chlorofluorocarbons (CFCs), hydrofluorocarbons (HFCs) and halocarbons. Each of these gases has the ability to absorb incoming solar radiation and outgoing terrestrial radiation at various wavelengths of the electromagnetic spectrum (Figure 25.2). Ozone is especially important in absorbing potentially harmful ultraviolet radiation in the upper atmosphere.

Figure 25.2 shows very clearly that if the atmospheric concentration of any or all of these gases changes, so too will the amount of solar radiation absorbed. If there is

an overall decrease and all other factors remain constant, there will be a net cooling of the earth's surface. Conversely, if there is an overall increase, there will be a net warming. This brings us to the principle of 'radiative forcing'. We have seen that the earth receives energy from the sun and loses energy to space mainly through infrared radiation. On average, there is a balance between energy inputs and outputs. Any factor that disturbs this balance is called 'radiative forcing'. The climate responds to such disturbance by changing (i.e., becoming warmer or cooler) until a new balance is achieved. A change towards a warmer state is popularly (and inaccurately) called the 'enhanced greenhouse effect'. (I say 'inaccurately' because the glass in a greenhouse allows the sun's heat to penetrate and warm the air inside the greenhouse, which slows the loss of outgoing long-wave radiation, but it does not itself absorb and store the heat in the same way that our atmosphere does.)

Let us now consider the geologically recent changes in certain atmospheric gases, such as carbon dioxide, nitrous oxide and methane (Figure 25.3). Trapped as air bubbles in the ice of Greenland and Antarctica is an archive of the mean atmospheric composition at the time the air within snow was finally sealed away from contact with the atmosphere as the snow turned to ice, a process that can take several centuries. The longest records come from ice cores collected by combined Russian and French teams from the site of Vostok in central Antarctica (Jouzel et al., 1997; Petit et al., 1981; Petit et al., 1990; Petit et al., 1999; Jouzel et al., 2007) and more recently by the EPICA team from EPICA Dome C (EPICA Community Members, 2004; EPICA Community Members, 2006; Lüthi et al., 2008). These ice cores span the past 800 ka. During that long interval of time, the atmospheric concentration of carbon dioxide ($p\text{CO}_2$) ranged between 280–300 parts per million by volume (ppmv) during interglacial maxima and 180–200 ppmv during glacial maxima (Jouzel et al., 2007; Lüthi et al., 2008). Corresponding methane values were about 800 ppbv during interglacial maxima and 400 ppbv during glacial maxima (Petit et al., 1999). As might be expected, temperatures were lowest on the ice caps during glacial maxima and highest during interglacial maxima. We also saw in Chapter 9 that concentrations of wind-blown dust from Patagonia and Australia were also highest during glacial times.

At the start of the Industrial Revolution (about 1750 AD), the carbon dioxide concentration during our present interglacial was about 280 ppmv (Table 25.1). From then on, it rose steadily and at an accelerating rate, reaching 379 ppmv in 2005 (Table 25.1) and 386 ppmv in 2009. By April 2013, the global $p\text{CO}_2$ level had reached 400 ppmv for the first time in nearly 1 million years, and it is continuing to rise at a rate of 2 ppmv/year (Raupach and Fraser, 2011). The same holds true for both methane and nitrous oxide concentrations, both of which are rising steadily (Table 25.1). Over the past 250 years, carbon dioxide has been the main contributor to radiative forcing (equivalent to about $+2.5 \text{ watts/m}^2$). Methane has an impact on radiative forcing that is about one-third that of carbon dioxide, and nitrous oxide (N_2O)'s impact

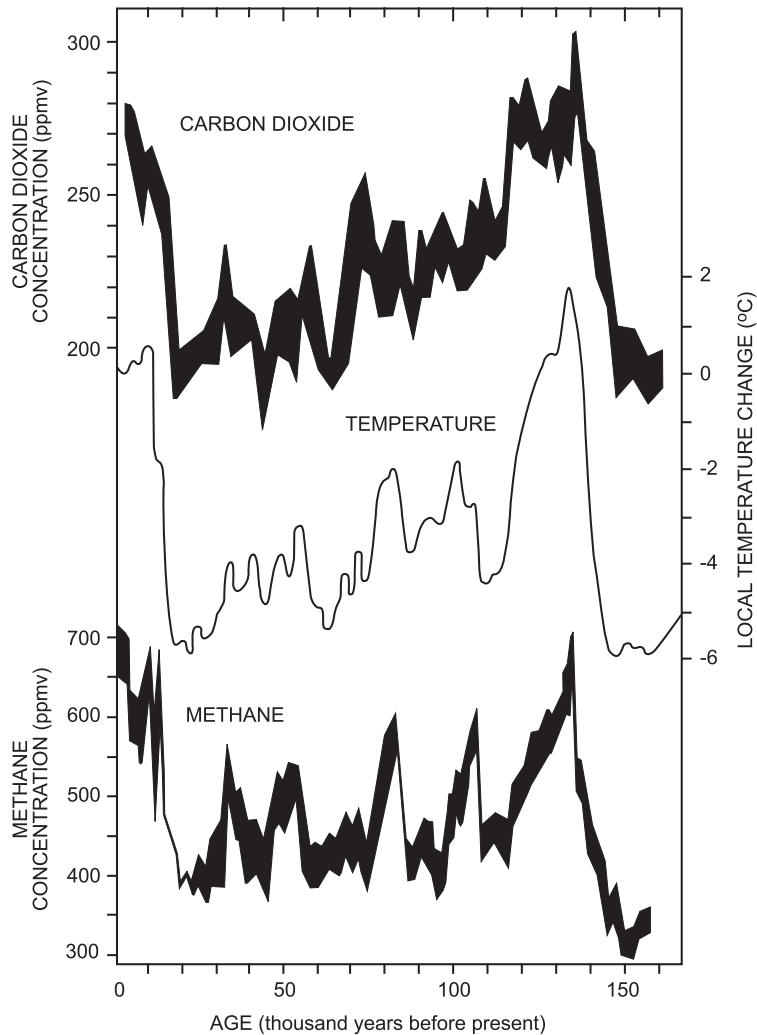


Figure 25.3. Changing concentrations of carbon dioxide and methane from air bubbles trapped in Antarctic ice during the past 150,000 years. (From Williams, 2001a, fig.1.8.)

Table 25.1. Summary of key greenhouse gases affected by human activities. (From IPCC, 2007a, and Houghton, 2009; ppmv is parts per million by volume; ppbv is parts per billion by volume.)

	Carbon dioxide (ppmv)	Methane (ppbv)	Nitrous oxide (ppbv)
Pre-industrial concentration	~280	715	270
Concentration in 2005	379	1774	319
Atmospheric lifetime (years)	50–200	12	114

is about one-tenth that of carbon dioxide. In the lower atmosphere, the increase in ozone concentration has a warming effect equivalent to or slightly greater than that of N_2O . Destruction of ozone in the upper atmosphere (stratosphere) as a result of ozone-depleting chlorofluorocarbons (CFCs) has a cooling effect at high elevations. Some gases and aerosols act to offset the warming effect resulting from the increase in CO_2 , CH_4 and N_2O , most notably sulphur dioxide (SO_2), which has a regional cooling effect.

Le Quéré et al. (2008) warned that there had been a 29 per cent increase in fossil fuel emissions between 2000 and 2008. Total carbon (C) emissions in 2010 amounted to about 9.28 billion metric tonnes, or gigatonnes ($9.28 \text{ Gt} = 9.28 \times 10^9 \text{ t C}$). (To convert tonnes of carbon to equivalent tonnes of carbon dioxide, simply multiply by 3.67). Burning fossil fuels and cement manufacture are the major sources of atmospheric carbon dioxide emissions (about 90 per cent) and amounted to roughly 8.35 Gt C, of which slightly more than half remains in the atmosphere (see Global Carbon Emissions, 2010, which updates the data in Canadell et al., 2007, and Le Quéré et al., 2008). Biomass burning and land use change, especially deforestation, are estimated to have contributed an additional 0.9 Gt C or so in 2010. During the last 600 years, there have been considerable variations in Southern Hemisphere biomass burning, with peaks centred around 1300 and 1900 AD and minima between 1300 and 1600 AD and after 1900 AD (Prentice, 2010; Wang et al., 2010). At the present time, there is a very real danger that increased carbon emissions from fires in areas of secondary forest regrowth in Amazonian forests and elsewhere may offset the carbon savings from curbing deforestation (Aragão and Shimabukuro, 2010). However, the work of Wang et al. (2010) has demonstrated that Southern Hemisphere biomass burning is at present lower than at any time in the previous 650 years, although, as Prentice (2010) has cautioned, this could change for the worse if global mean temperatures increase by several degrees. Apart from the carbon stored in the atmosphere, the sinks for carbon were estimated at about 2.23 Gt C for the oceans in 2010 and about 2.41 Gt C/year for the biosphere.

Major sources for methane are more varied than they are for carbon, and they include fossil fuel mining and distribution, livestock, landfills, rice paddies, wetlands and biomass burning, in broadly similar proportions. There is considerable concern over current permafrost thawing in high northern latitudes, which could release large additional supplies of methane hitherto locked up in frozen swamps (Schuur et al., 2011). Ruddiman and Thomson (2001) have suggested that the increase in methane (CH_4) emissions evident in Northern Hemisphere ice cores from about 5,000 years ago onwards was a result of inefficient rice cultivation in flooded areas with abundant CH_4 -producing weeds. They based this conclusion on the fact that the 100 ppbv increase in atmospheric CH_4 after 5 ka followed a pattern not seen previously in the orbitally driven ice core record. This attractive hypothesis is still in need of rigorous testing.

25.4 Current global and regional trends in temperature and precipitation

Global mean land temperatures have increased by roughly 0.7°C during the past 100 years, and sea surface temperatures have increased by 0.9°C since 1900 (Raupach and Fraser, 2011). Between 1961 and 2008, the upper few hundred metres of the oceans absorbed about 1×10^{22} Joules of energy, which has also helped slow the rate of warming in the lower atmosphere (Braganza and Church, 2011). The rise in temperature on land has fluctuated, but the overall trend is upwards. The oceans have also become progressively warmer during that time, resulting in thermal expansion and a rise in global sea level at a rate that in 2010 amounted to 3.2 mm/year (Braganza and Church, 2011). Between 1880 and 2010, mean global sea level rose about 20 cm, about half of which is attributable to thermal expansion of the warmer ocean surface and about half of which is the result of the melting of ice stored in mountain glaciers and ice caps, including Greenland and Antarctica. We saw in Chapter 13 that the mountain glaciers in Africa have been retreating rapidly over the past fifty years, and this is true of the majority of mountain glaciers in both hemispheres. Global sea level rise is likely to continue as the surface of the oceans continues to warm and ice caps and glaciers continue to melt. Given that more than 150 million people today live within one metre of mean sea level, this is a sobering trend. Should the Greenland ice cap melt entirely, as it has in the geologically recent past, it is estimated that global sea level would rise about 7 m, and if the entire West Antarctic ice cap melted, there would be an additional 5 m rise (Braganza and Church, 2011). During the peak of the last interglacial 125,000 years ago, mean sea level is estimated to have been 6–9 m higher than it is today, reflecting meltwater contributions from both Greenland and West Antarctica (Dutton and Lambeck, 2012).

The synchronous increase in both land and sea surface temperatures during the last hundred years points to a global cause, with the warming of the lower atmosphere as a result of the enhanced greenhouse effect being the most probable cause. Global climate models that only take into account natural forcing factors, such as volcanic eruptions, solar variability and ENSO events, fail to reproduce the monitored changes in global temperature. When the models include both natural and anthropogenic forcing factors, the match between model outputs and observations is much improved (IPCC, 2007a), suggesting that anthropogenic emissions of the greenhouse gases such as CO₂, CH₄ and N₂O are indeed a significant factor in the global warming trend.

Warming of the land due to an increase in the carbon dioxide content of the lower atmosphere is entirely consistent with basic atmospheric physics, as Svante Arrhenius demonstrated more than a hundred years ago (Arrhenius, 1896). Building on the solid experimental work of John Tyndall and on Joseph Fourier's identification of '*l'effet de serre*', or the greenhouse effect, Arrhenius calculated that a reduction in the atmospheric CO₂ concentration to 0.67 of the late nineteenth-century value of about 300 ppmv would lower the mean surface temperature by 3°C, while increases

to 1.5, 2, 2.5 and 3 of that value would lead to increases of 3–4°C, 5–6°C, 7–8°C and 8–9°C, respectively (op. cit., 1896, p. 266, table VII).

Part of his motivation for carrying out these very laborious calculations was the dissatisfaction he felt about contemporary attempts to explain the recent ice ages. Arrhenius (1896, p. 274) was especially scathing about James Croll's astronomical theory of ice ages (see [Chapter 3](#)), quoting with approval Luigi de Marchi's conclusion (translated): 'Now I think I may conclude that from the point of view of climatology and meteorology, in the present state of these sciences, the hypothesis of Croll seems to be wholly untenable as well in its principles as in its consequences' (de Marchi, 1895, p. 166). Arrhenius (1896, p. 274) was more inclined to favour de Marchi's view that a change in atmospheric transparency caused by changes in CO₂ concentration was a more likely explanation of the global cooling and warming associated with the ice ages. He had earlier commented that the temperature of the Arctic at an unspecified time in the Tertiary had been 8–9°C warmer than today on the basis of the fossil flora, and he considered that long-term changes in the atmospheric CO₂ concentration may have been responsible.

It could be argued that the monitored temperature record from North America and north-west Europe is flawed because of the urban heat island effect and that the warming trend evident over the past hundred years is more a function of urbanisation than of any global trend. One obvious counter-argument is that the synchronous warming of the oceans has nothing to do with urbanisation. Another is that non-urban areas show a similar warming trend. Consider, for example, Australia, where more than 90 per cent of the population live in five widely scattered state capital cities (Perth, Adelaide, Melbourne, Sydney and Brisbane). The many small towns and rural settlements across this vast dry and sparsely inhabited continent all show a warming trend. In the 100 years between 1910 and 2009, the mean surface temperature across Australia increased by 1°C (Braganza and Church, 2011). The decade of 2000–2009 in Australia was the warmest on record. Most of the continent has become warmer over the last fifty years, with the lowest rate of warming in the north-west also being associated with an increase in mean annual precipitation since 1960. These two phenomena may be related. As a general rule, years of highest rainfall are years of lowest mean temperature, and vice versa. The greatest warming has been in the spring, and there has been an increase in the frequency of warm weather, with the number of hot days per year increasing progressively from 1960 to 2010 (Braganza and Church, 2011).

Establishing long-term trends in precipitation is not easy, especially in a continent like Australia, which has the most variable rainfall of any continent or desert region, particularly in the eastern half of the continent, which is subject to the influence of ENSO events (Kuhnel et al., 1990; Peel et al., 2002). There has been a 15 per cent decrease in winter precipitation in the south-west and south-east of Australia since the mid-1970s, which is larger than the natural variability. One outcome has been a 60 per cent decline in annual stream flow in the lower Murray-Darling Basin and in the

south-west (Braganza and Church, 2011). As a result of these trends towards reduced winter rainfall, the cities of Perth in the south-west and Adelaide in the south have both invested in expensive desalinisation plants in order to secure their drinking water supplies.

Concern over possible adverse impacts of future changes in climate has stimulated research into historic trends in precipitation and evaporation (Suppiah et al., 2006; Pittock, 2009; Cleugh et al., 2011). The Australian Bureau of Meteorology has produced a series of maps showing changing regional patterns of mean annual precipitation between 1900 and 2009 (Holper, 2010). Summer rainfall in the tropical north-west of the continent has increased progressively over the past hundred years and especially over the past fifty years, while winter rainfall in the south-west, south and south-east of the continent has decreased, especially since the mid-1970s. Although these changes are more or less within the realm of natural rainfall variability within Australia, they will soon lie outside those limits if current trends persist. We can say with confidence that although the climatic changes evident across Australia do not necessarily *demand* anthropogenic forcing, they are certainly *consistent* with such forcing.

25.5 Possible impacts of future changes in climate on the desert world

Predicting the impacts of possible future changes in climate is based on three general approaches, none of which is entirely satisfactory. One involves the use of global atmospheric circulation models, which have become increasingly complex in the last few decades. Another involves extrapolating from what we can reconstruct of past climates. The third involves using current climatic trends.

A fundamental problem with all current climate models is their inability to make credible predictions about changes in cloud cover and water vapour. Clouds are highly dynamic and notoriously hard to model, and so are either omitted from climate models or included in a highly schematic manner. Water vapour is the most potent of the greenhouse gases, and is also very hard to model realistically. As a consequence, current models are often more akin to sensitivity tests than otherwise when it comes to predicting changes in regional and local precipitation, and do not really provide us with watertight predictions that natural resource managers can use with confidence. It is hard to see how this difficulty can be easily overcome. The unresolved cloud problem also accounts for why existing models vary widely in predicting future changes in the spatial and temporal distribution of precipitation but are all reasonably consistent in predicting likely changes in temperature associated with various levels of greenhouse gas emissions, since this is based on better-known aspects of atmospheric physics (IPCC, 2007a; Houghton, 2009).

Extrapolating from inferred past climatic changes is equally fraught with uncertainty, not least in terms of knowing whether present-day boundary conditions are the same as they were in the past. These boundary conditions include the distribution of

land, sea and ice, ocean currents, albedo, plant cover and atmospheric composition. The further back in time we go, the more the boundary conditions differ from those of today. Perhaps the most useful aspect of studies of past climate from the point of view of providing insights into possible future impacts is in providing a spectrum of natural climatic variability against which to assess current trends.

Using climatic trends based on monitored observations of temperature, precipitation and evaporation may at first seem to be the most reliable of these three approaches to assessing the possible impacts of climate change. When confronted with extreme events such as prolonged droughts and widespread floods, which appear to be beyond the limits of natural variability, there is a natural tendency for climate scientists and natural resource managers to invoke ‘global warming’ or ‘global climate change’ as the culprit. This raises the question of what we mean by ‘natural variability’. There is no objective or absolute measure of this concept, because it depends entirely on the spatial and temporal scale adopted. It would be useful if those who use the term ‘natural variability’ were to define it relative to different time scales, which they seldom, if ever, do. One lesson from the past is that we can match modern extreme events simply by examining the Holocene record of past climatic and hydrologic fluctuations. That being the case, all we can safely say when dealing with modern extreme events is that they are consistent with present global climatic warming trends but are not necessarily caused by them.

Another wild card involved in assessing future climatic impacts in the drier regions of the world concerns possible future changes in ENSO events, which are presently responsible for substantial interdecadal and interannual variability in precipitation (Chapter 23). We know the areas that are most influenced by ENSO events today and over the past five hundred years and more, and can probably safely assume that this spatial pattern is unlikely to change very much in coming decades (Collins et al., 2010). However, we cannot predict with any confidence whether progressively warmer sea surface temperatures will lead to more frequent or more severe ENSO events than in the historic past (IPCC, 2007a). Some authors have proposed that change is likely, while others are less convinced. Vecchi et al. (2006) considered that in the future, the atmospheric circulation in the tropical Pacific would weaken, leading to weaker ENSO events, and Power and Smith (2007) found that the average June–December values of the Southern Oscillation Index between 1977 and 2006 were the lowest on record, indicating a weakening of the Walker Circulation (Chapter 22) at a time when the tropical sea surface temperatures were the highest on record. An important conclusion from this work was the need to take into account global warming when assessing links between ENSO events and future climatic variations across the globe.

We saw at the start of this chapter that there have been claims of an increase in the magnitude and frequency of extreme events such as floods and droughts as global temperatures on land and sea become progressively warmer. Here again, a measure of healthy scepticism is required. Consider the question of extreme droughts. In North

America, the most widely used index to assess drought severity is that devised by Palmer (1965), which is known appropriately as the Palmer Drought Severity Index (PDSI) and was discussed in detail in [Chapter 23](#). The PDSI provides a rapid means of assessing soil moisture deficits in different regions across the United States and hence of allowing relative drought severity to be compared. During the 1700–1978 period, based on analysis of 425 tree-ring chronologies, the ‘Dust Bowl’ drought of the 1930s proved to have been the most severe (Cook et al., 1999). Cook et al. (2007) have expanded on their earlier work and have developed a very useful North American Drought Atlas.

Using a similar methodology based on the PDSI and a network of 327 tree-ring chronologies, Cook et al. (2010) produced the Monsoon Asia Drought Atlas. The results showed four intervals of severe and widespread drought during the last thousand years, centred on 1638–1641, 1756–1768, 1790–1796 and 1876–1878. It is interesting that these were all times when the volcanic dust veil index compiled by H. H. Lamb showed peak values (Lamb, 1970; Lamb, 1972; Lamb, 1977), and they were therefore related to cooler-than-average weather in the Northern Hemisphere mid-latitudes.

It thus seems that where there is a detailed tree-ring chronology, the PDSI can serve as a useful tool for comparing the relative severity of different historic droughts. Whether the PDSI can be used to check recent trends in drought magnitude and frequency, however, is a matter for debate. Sheffield et al. (2012) have critically analysed the methodology used to construct the PDSI. They pointed out that the PDSI uses temperature data to calculate potential evaporation (E_{pot}) and ignores other key controlling variables, such as near-surface radiation, wind speed and humidity (see [Chapter 23](#) for details). If global temperatures are increasing, as observations indicate to be the case, a formula based on temperature will necessarily show an increase in drought frequency. Sheffield et al. (2012) used a physically based equation to determine potential evaporation and found very little evidence of any change in global drought frequency over the past sixty years, which is also consistent with assessments of droughts in North America (Karl and Heim, 1990; Idso and Balling, 1992; Soulé, 1993). These results flatly contradict the IPCC (2007a) assertion (based on evidence obtained from using the PDSI) that: ‘More intense and longer droughts have been observed over wider areas since the 1970s, particularly in the tropics and subtropics. Increased drying linked with higher temperatures and decreased precipitation has contributed to changes in drought’. The 2012 IPCC report on extreme events displays commendable caution in discussing drought trends, commenting on the earlier undue reliance by the IPCC on the PDSI resulting in possible overestimation of the increase in regional and global droughts (Seneviratne et al., 2012).

The crux of the problem in predicting likely future climatic impacts revolves around the degree to which human activities have altered and are likely to alter global and regional climates, leading us into hitherto uncharted waters. However, we can draw on the geological insights of James Hutton (1795) and Charles Lyell (1830–1833)

and make use of present-day extreme events to reconstruct past geographies in ways that were not possible using global atmospheric circulation models. For example, the extreme rains during the 1999 wet season in arid central Sudan filled the hollows in the extensive dune complex immediately east of the lower White Nile and recreated the landscape of early Holocene times (Williams and Nottage, 2006). At this time, Mesolithic hunter-gatherers occupied seasonal camps on top of dunes flanked by flooded swales, and ventured forth to collect *Pila wernei* shells for food, to harpoon hippos and to fish for Nile perch (Clark, 1989; Williams and Nottage, 2006).

Keeping in mind all of these caveats, we are now in a position to attempt some cautious predictions about some of the impacts of possible future climatic changes. In the Southern Hemisphere, the southward displacement of the Antarctic Convergence Zone appears to be associated with a concomitant southward displacement of the air masses that bring winter rainfall to south-west Australia, South Africa and South America. If this trend persists, over the next few decades, we would expect a further reduction in winter precipitation and in stream flow in all of these regions, consistent with trends over the past forty years. In contrast, the tropical summer rainfall regions in the Southern Hemisphere have been receiving higher amounts of precipitation, leading to episodically widespread flooding in north-east Australia, Mozambique and north-east Brazil. The net increase in temperature over the past hundred years has seen an increase in spring snow-melt and river discharge in south-east Australia and the Andes and a reduction in the size of ice caps and mountain glaciers both in the Andes and in equatorial Africa (Ruwenzori, Kilimanjaro and Mount Kenya). The increase in the number of very hot days, when combined with high fuel loads, drought and strong winds, has led and will lead to extreme bushfire risk, for which communities will need to plan. As far as the southern deserts and their margins are concerned, the tropical northern margins will probably become less arid and the southern margins will become more arid, with a possible increase in extreme episodes of drought and heavy rainfall. In addition, the increase in tropical sea surface temperatures will likely to lead an increase in the incidence of tropical cyclones and hence to higher rainfall in those regions most prone to cyclones today, such as the arid north-west of Australia.

In the Northern Hemisphere, two dramatic changes are already very evident. One is the steady reduction in the area of Arctic sea ice, which has reduced the albedo in this region and has thus allowed more solar radiation to reach the surface, causing further melting of snow and ice – an excellent example of a positive feedback process in action. The other is the thawing of vast areas of permafrost in northern Eurasia, leading to release of methane into the atmosphere, which will in turn accentuate high-latitude warming. Mountain glaciers and ice caps are in steady recession across North America and Eurasia. There is concern that as run-off from snow-melt eventually diminishes in California and the American Southwest, there will be a reduction in river discharge and in groundwater recharge. Higher temperatures and drier soils have already resulted in severe bushfires in California and in the Mediterranean countries of Europe, notably

Portugal, Spain, France, Italy and Greece. There has also been a significant increase in deaths from heat stress among the poor and elderly in Europe and North America. Another major issue is a change in the incidence of climate-related diseases, such as malaria, dengue fever and meningitis, with expansion of these diseases into areas that were previously exempt, as in the highlands of Ethiopia and Kenya. On a more positive note, if the early Holocene can be used as a guide to what might happen as sea surface temperatures become warmer, there is likely to be a strengthening of the summer monsoon across Asia and Africa, leading to a longer summer wet season with more intense rains. This should lead to an increase in plant cover along the southern margins of the tropical deserts, such as the Sahara and Thar deserts, but it will also depend on human pressure in the form of deforestation and overgrazing, as discussed in [Chapter 24](#). When confronted with extreme climatic events in the past, human societies have responded in one of three ways: they migrated (Clark, 1980), adapted (Macklin et al., 2013) or became extinct (Diamond, 2005). Those societies with flexible patterns of resource use were the ones that survived best. [Chapter 26](#) outlines some key prerequisites for achieving sustainable use of the natural resources of our desert regions. In a world of uncertainty, conservative use of water, energy, plants and animals is essential, as are flexible forms of land use. The most adaptable communities are the ones that will fare best in coping with the uncertainty inherent in future climatic change.

25.6 Conclusion

The earth's climate is governed by energy from the sun. About half of all incoming short-wave solar radiation reaches the surface of the earth and is then emitted back to space as outgoing long-wave (infrared) terrestrial radiation. Roughly half of all solar radiation is received at the earth's surface, with some reflected back to space and the rest absorbed by the atmosphere. If there were no atmosphere around the earth, the average temperature at the surface of the earth would be -18°C rather than the present $+14^{\circ}\text{C}$. This is because certain gases in the atmosphere absorb infrared radiation and cause the lower atmosphere to become warmer. Such gases include water vapour, carbon dioxide, methane and nitrous oxide. These gases are popularly known as greenhouse gases.

At any given time, there is a balance between energy inputs to the earth's surface from short-wave solar radiation and energy losses from the surface of the earth in the form of long-wave terrestrial radiation. Any factor that disturbs this equilibrium will lead to either cooling or warming at the earth's surface, depending on whether it causes a net increase or decrease in the solar energy that reaches the surface of the earth. For example, an increase in sulphur dioxide can lead to localised cooling, while an increase in any of the greenhouse gases has the reverse effect. At the present time, there is a steady increase in the atmospheric concentration of carbon dioxide,

methane and nitrous oxide relative to the preceding one million years. The increase began around 1750 AD (and possibly earlier in the case of methane), as the Industrial Revolution gained momentum, resulting in the increasing burning of coal and, later, oil, as well as widespread forest clearance in Europe and, more recently, in the tropics. The increase in atmospheric methane concentration may have begun about 5,000 years ago, as inefficient rice cultivation began in Asia. Before the Industrial Revolution, the maximum concentration of atmospheric carbon dioxide (pCO_2) amounted to about 280–300 ppmv. In April 2013, it was 400 ppmv, and is increasing at an accelerating rate. There have been fluctuations in pCO_2 throughout the Cenozoic, but what seems to be new is the speed with which the concentration of the various greenhouse gases is increasing, which appears to be without equivalent in geologically recent times.

An expected consequence of the increase in the concentration of greenhouse-enhancing gases in the lower atmosphere has been a net warming of the surface of both land and sea amounting to about 1°C on average, with greater warming evident in higher latitudes. This warming is evident in rising sea levels and in the widespread melting of ice caps and mountain glaciers. About half of the sea level rise is estimated to be from melting ice and about half from thermal expansion of the warmer ocean surface.

Efforts to predict how the earth's climate might respond in the near future draw on a combination of global atmospheric circulation models, historical archives, meteorological observations and insights from the study of past climates. There is enormous uncertainty over possible future changes in precipitation around the globe. Present-day trends suggest that winter rainfall may continue to decrease in those regions in the Southern Hemisphere that experience a Mediterranean type of climate. Conversely, summer rainfall is likely to increase along the margins of the tropical deserts, although precipitation incidence may become more erratic, with more frequent floods and droughts. The adverse impacts of possible future changes in climate are likely to be greatest among the poorest communities that are already subject to erratic rainfall regimes and widespread land degradation. Many such communities now live along the desert margins and in the dry subhumid tropics, and they have a long history of adapting to climatic vicissitudes. They will need to display similar levels of resilience and adaptable strategies of grazing and farming in the foreseeable future. Good governance, food security, access to clean drinking water and health-care facilities are necessary to ensure success. However, it is not only the poorer rural societies who will be most vulnerable. The more advanced societies that are highly dependent on technology and major infrastructure will also find it hard to adapt, especially large cities, where good drinking water is already in scarce supply.

Towards sustainable use of deserts

He prayeth best, who loveth best
All things both great and small
Samuel Taylor Coleridge (1772–1834)
The Rime of the Ancient Mariner (1798)

26.1 Introduction

In the opening chapters of this volume, we saw that the distribution of our deserts is governed by long-established tectonic, topographic and climatic factors and that the onset of climatic desiccation in all of the world's deserts began many millions of years ago and had nothing to do with any recent human influence. Evidence from pollen analysis, molecular genetics and sediment geochemistry, discussed in later chapters of the volume, also confirms the antiquity of the deserts. However, within and beyond the confines of the present-day deserts, there is abundant and well-dated evidence that past climates in what are now arid and semi-arid environments have frequently been wetter than they are today. This evidence is derived from defunct river systems and fossil lakes within the deserts, with their rich stores of aquatic fossils, as well as from cave deposits and fossil soils. We also saw that the response of rivers, lakes and glaciers to climatic change is often quite variable, so we should avoid relying on only one source of information.

The fossil remains of the first bipedal hominids date back more than 5 million years in what is today the arid Afar Rift of Ethiopia and the desert region of northern Chad. The rich vertebrate fauna from these two areas indicates that the climate at that time was somewhat less arid than it is today. These wetter intervals alternated with more arid phases, during which vast deserts like the Sahara were even more extensive than at present and desert dunes were active well beyond their existing limits. At these times, great plumes of desert dust were blown out to sea, often as far as the Amazon Basin, Greenland and Antarctica. Some of this dust remained on land to form

extensive mantles of silt, of which the deep loess deposits in north-central China are the thickest and best-known.

The well-preserved sedimentary and microfossil evidence from deep-sea cores complements the more fragmentary terrestrial record from the deserts and their margins and shows that around 2.5 million years ago, the earth's climate became colder, drier and more variable than it had been earlier. This was also the time when the first stone tools appear in the African archaeological record, a precursor to the progressively more intense interactions between prehistoric humans and their environments, including the eventual use of fire for clearing vegetation and modifying the plant cover.

As we come closer to the present, the desert record of past climatic changes becomes ever more detailed (Table 26.1), and with the discovery of writing, the archival record of past floods and droughts can be compared with that from tree rings and from precisely dated flowstone deposits in desert caves. This information reveals that climatic change in deserts has operated at a variety of scales in time and space, with close links between global sea surface temperature anomalies and the incidence of historic floods and droughts. Volcanic eruptions have also played a role, particularly those by explosive volcanoes located in tropical latitudes. Historic eruptions have been associated with short-term global cooling, the effect lasting only one or two years, but there is some indication that such eruptions also helped alter the regional water balance and led to reduced rainfall.

One question that has often generated more heat than light is the thorny problem of desertification and, in particular, the relative importance of climatic fluctuations and human interventions in promoting desertification processes. Part of the problem associated with the present discourse on desertification stems from conflicting definitions. At present, desertification is defined as 'land degradation in arid, semi-arid and dry subhumid areas *resulting from various factors including climatic variations and human activities*' (UNEP, 1992b, italics added). Because desertification reflects a state or condition of the land, notably, degraded land that is less able to support life than it was previously, it might have been wiser to leave the question of causes out of the definition and to omit that portion of the definition shown in italics. To say that desertification is caused by various factors is stating the obvious, and it is equally obvious that both climate and humans play a role. There are many different factors that can lead to land degradation, both directly and indirectly (see Chapter 24, Figures 24.1 and 24.2). Ultimately, many of them are bound up with social, economic and political factors that have for too long been ignored in the current debates.

The exponential increase in world population has led to increasing demand for natural resources and increasing pressure on the land (Rockström et al., 2009). In the drier parts of the world, this has meant drawing on water from deep-seated aquifers that are no longer being recharged at a rate comparable to the rate of withdrawal – in effect, mining the groundwater. More people and more animals on the land have led

Table 26.1. *Summary of global and regional environmental and climatic changes discussed in this volume from 250,000 years ago to present*

I. 250–25 ka

(Interval includes two glacial-interglacial cycles, each ~100 ka long)

- High interglacial sea levels at ~210 and ~125 ka
- Warmer interglacial sea surface temperatures, high rates of evaporation from the oceans, and wetter climates in the intertropical zone
- High lake levels in Africa and Australia at ~130–110, ~100–80, ~65 and ~40 ka
- Low glacial sea levels at ~150 ka; large ice caps over North America and north-west Europe; glacier advances in South America
- 74 ka eruption of Toba volcano, possibly followed by several decades of intense global cooling and several centuries of sustained reduction in precipitation in Asia and perhaps Africa (controversial)
- Prolonged wet phase at ~125 ka; Sahara and peninsular Arabia studded in lakes, with ITCZ displaced up to 500 km further north over North Africa during Northern Hemisphere (NH) summer
- Stronger summer monsoon in Asia during last two interglacial phases
- Sapropel accumulation in East Mediterranean Sea at 217 ka (S8), 195 ka (S7), 172 ka (S6), 124 ka (S5), 102 ka (S4), 81 ka (S3) and 55 ka (S2), broadly coincident with phases of very high Nile discharge

II. 25–17 ka

(Interval includes the Last Glacial Maximum (LGM: 21 ± 2 ka))

- Maximum global ice volume and LGM sea levels ~120 m lower than today
- Small ice caps and more extensive mountain glaciers in Africa, South America, Australia and Asia
- Cooler sea surface temperatures and reduced evaporation from the oceans, with drier conditions in the intertropical zone
- Weaker summer monsoon in Asia, North Africa and northern Australia
- Lakes low or dry in Africa, India, China and Australia during the LGM
- Desert dunes active in Africa, Asia and Australia; widespread deposition of desert dust/loess in China, India, South America, Australia and Europe
- High lakes levels in drier parts of North America

III. 17–5 ka

- Retreat of ice caps and mountain glaciers in both hemispheres and rapid sea level rise from 17 ka onwards
- Sea surface temperatures higher once more, with greater evaporation from the oceans and wetter climates in the intertropical zone
- Stronger summer monsoon in both hemispheres
- High postglacial lake levels in Africa, Asia and Australia
- Desert dunes vegetated and stable along desert margins; supply of desert dust/loess much reduced
- Stronger summer monsoon; ITCZ extends ~500 km further north than today during the NH summer
- The Sahara is once more studded in sporadic lakes and supports a human population of Mesolithic hunter-fisher-gatherers and later Neolithic pastoralists

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- Rivers of central India cut down ~30 m into their former floodplains at or before ~8 ka
 - Ice-rafted debris (IRD) events in North Atlantic at 11, 10.3, 9.4, 8.1 and 5.9 ka
 - Donggwe Cave speleothem record (southern China) shows strong Asia Monsoon (AM) at 9–7 ka and weak AM events at 8.3, 7.2, 6.3 and 5.5 ka
 - Oman speleothems show reduced rainfall at 9.5, 9, 8.3, 7.4 and 6.3 ka

IV. 5–0 ka

- Climates in the intertropical zone become less humid and more seasonal
 - Summer monsoon still strong but less vigorous than in the previous phase and its spatial domain probably somewhat reduced
 - Lakes dry out in arid and semi-arid areas, including the Sahara, which becomes abandoned by Neolithic pastoralists who move south into West Africa or east into the Nile Valley
 - Weaker summer monsoon; ITCZ retreats ~500 km to the south in North Africa during the NH summer
 - Sporadic glacier advances culminating in the most recent Little Ice Age
 - Increase in the frequency of El Niño-Southern Oscillation (ENSO) events, leading to more frequent extreme floods and droughts in the Americas, Africa, Australia and Asia; interaction between ENSO and summer monsoon leads to more variable rainfall regime in both hemispheres in regions influenced by them and to highly seasonal or very variable river flow regimes
 - Impact of human activities on global climate (and on river basins) may have begun with the advent of agriculture and has increased ever since, not always to the advantage of either
 - IRD events in North Atlantic at 4.2, 2.8 and 1.4 ka
 - Donggwe Cave speleothem record (southern China) shows weak Asia Monsoon events at 4.5, 2.7, 1.6 and 0.5 ka
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to the overgrazing of rangelands in every continent – to the detriment of both the land and the humans dependent on it. Accelerated clearing of the native vegetation has denuded the land, making it more and more susceptible to soil loss by wind and water. In certain regions, such as semi-arid southern Australia (Powell, 1976; Powell, 1988), such clearing has caused the regional groundwater-tables to rise, bringing dissolved salts to the surface and leading to what has been called dryland salinization. Attempts to grow crops in desert regions using irrigation water from rivers originating well outside the deserts, such as the Indus River in Pakistan, the Nile in Egypt and Sudan, and the Amur Darya and Syr Darya in Uzbekistan, have often led to waterlogging and salt accumulation in soils, as well as problems downstream, including the silting up of rivers and drying out of desert lakes and inland seas, such as the Aral Sea in central Asia and a number of once large lakes in Iran.

Given the magnitude of these problems of land degradation, groundwater depletion, and loss of plant cover and biodiversity in the desert world, it is important that we identify a template or a set of guidelines for both present and future sustainable use of the deserts and their extensive semi-arid and dry subhumid margins. Two principles of

sustainable land use in deserts, based on the first and second laws of thermodynamics, are therefore offered in this final chapter, together with two other key preconditions, one based on social justice, the other based on the vital role of plants in minimising or avoiding land degradation.

26.2 Prerequisites for achieving sustainable use of deserts

We have discussed a number of examples of land degradation in the drier regions of the world that were caused by human activities and climatic variations ([Chapter 24](#)) and have also considered how possible future changes in climate may have an impact upon the desert world ([Chapter 25](#)). It is now appropriate to consider some fundamental principles of sustainable land use that should underpin any human interactions with these ecologically fragile regions.

We begin with the proposition that matter and energy are interchangeable, a proposition expressed in 1905 by Einstein in concise mathematical form in his theory of special relativity. In classical physics, the first law of thermodynamics states that energy can neither be created nor destroyed, so a system can lose or gain energy only to the extent to which it passes it to, or takes it from, its immediate surroundings. (The concept of stream power, discussed in [Chapter 10](#), nicely illustrates this point.) The second law of thermodynamics states that all natural processes lead to an increase in the entropy of the system concerned. Entropy is a measure of the unavailable energy in a thermodynamic system and reflects the state of disorder of that system. (Before beginning work on this volume, my books and papers were neatly arranged in what I consider good order; by the end of the writing process, they were scattered in disarray: entropy within the system had increased.) We are now in a position to reformulate the first and second laws of thermodynamics in the form of guiding principles for sustainable use of drylands, following the path laid out by the polymath cancer specialist Dr Karl-Henrik Robèrt ([1992](#)) in his book entitled the *Nödvändiga Steget* (Swedish for ‘natural step’). The fourth prerequisite is based on an appeal to social justice, and it is explained in number 4.

1. From the outset, we need to recognise that the only source of a *net* increase in primary productivity on this earth is from solar energy acting through photosynthesis to increase plant matter. All else is simply applying energy to recycling earth materials (such as coal and iron) into other forms of matter (such as steel and carbon dioxide). For this process of photosynthesis to operate effectively, biomass needs to be protected and increased and not, as is often the case at present, wantonly destroyed.
2. From classical thermodynamics, we can deduce that we should not *systematically* remove materials from any natural or humanly modified system at a rate faster than that at which they can be replaced. (For instance, soils need to be maintained or replenished at about the same rate at which they develop through rock or sediment weathering and biological activity.)

3. Likewise, a second deduction is that we should not *systematically* add materials to a natural or humanly modified system at a rate faster than that at which the system can absorb and recycle those materials. (Excessive use of pesticides, herbicides and fertilizers is one example of breaching this precondition, as is the general problem of chemical pollution of soils, water and air.)
4. The final prerequisite, based on simple human justice, is that there should be fair and effective access to natural resources by all people of the globe. (Access to clean, safe drinking water is an obvious and pressing example). We will now consider how each of these four conditions, or prerequisites, can be put into practice in order to achieve sustainable use of the natural resources of the desert world.

26.3 Safeguarding desert ecosystems and retaining plant cover

The Greek philosopher Empedocles (ca. 495–435 BC) believed that life on earth was made up of four ultimate elements, namely, earth, air, fire and water, an idea that reflected far older Babylonian concepts. Thinkers from every great civilization and religion have speculated about the origin and nature of life on earth (Glacken, 1973), and these ideas have been greatly refined over the last few centuries. However, it is only comparatively recently that we have come to realise the crucial importance of biological diversity (or biodiversity) for maintaining the health of soil, air and water on this planet. The seasonal fluctuations in the global atmospheric concentration of carbon dioxide are a direct reflection of reduced photosynthesis in northern mid- and high latitudes in winter and the renewal of plant growth in the boreal (northern) spring and summer. The removal of forest and burning of the savannas means less oxygen input to the atmosphere and an increase in emissions of carbon dioxide and carbon particles. Biomass burning can also lead to increased run-off and erosion (Dube, 2009). Common sense therefore tells us that any removal of the plant cover in one place should be balanced by its replenishment in another. Failure to achieve such a balance means a reduction in global biomass and a change in atmospheric composition. Consequences that are more obvious because they are more visible include the dust storms and mobile dunes that are engendered by the stripping of the former stabilising plant cover. Soil nutrients are generally concentrated in the organic rich top-soil. The presence of soil humus also promotes soil structural stability and higher rates of infiltration, thereby minimising erosion from surface run-off or overland flow.

People who have spent their lives in deserts are well aware of the importance of maintaining an appropriate plant cover. The Kababish camel herders of northern Sudan may not be aware that the earth is an oblate spheroid, but they are certainly aware that their survival depends on only taking what is needed for their animals, and being selective at that. Charles Doughty in his wanderings through ‘Arabia Deserta’ in the late 1870s mentions the scorn exhibited by the Bedouin nomad children when he collected shrubs valued as camel fodder to put on the fire (Doughty, 1956, p. 79).

A century later, the geographer A.T. Grove pointed out that ‘nomadic flexibility is an advantage to people living in fluctuating, marginal environments, and nomadic mobility allows good use to be made of variable grazing’ (Grove, 1974, p. 151). He went on to observe that unfortunately the traditional systems of pastoral nomads ‘do not fit neatly into the framework of a modern state’ (op. cit., p. 151). The problem is exacerbated in times of severe drought when nomadic pastoralists move into land that has been cultivated by sedentary farmers during the dry season, especially when traditional modes of conflict resolution have been taken over by a remote central government. The roots of the current violence in Darfur in arid western Sudan stem from these causes. Friction between herder and cultivator is nothing new, as shown by the recurrent nineteenth-century conflicts between cowboys and sodbusters in the drylands of the North American Southwest.

A tricky problem in many degraded rangelands is how to separate out the effects of climatic fluctuations such as drought from those of overgrazing and indirect human modification of the vegetation communities (Nicholson, 2011). One useful approach to this complex issue, tested in both Australia and Botswana, is to monitor changes in land cover along selected transects, starting near the main watering points where overgrazing pressure is likely to be greatest, and to see whether plant recovery following the end of a drought returns to the same condition as in areas deemed to be the least degraded (Pickup, 1998; Dube and Pickup, 2001). In an effort to assess plant cover across six types of terrain common in semi-arid south-east Botswana, Dube (1998) used different wave bands of SPOT and Landsat imagery but found that the variance in vegetation within a given landscape exceeded the variance in vegetation cover between different landscapes. She later developed a more precise method of estimating plant cover, in this instance woody plant density in communal and leasehold land tenure systems in semi-arid north-west Botswana, using aerial photographs and a carefully devised set of protocols (Dube, 2008). These studies show the difficulties involved in gauging changes in plant cover, and also highlight the importance of taking into account differences in spatial scale.

Although there is far less in the way of both biomass and biodiversity in the desert world than in the tropical rainforests, the harsh conditions in deserts have led to a wide variety of adaptations to hot, dry environments that may prove to have great survival value should severe and prolonged droughts become more common in the future (see Chapter 25). In public discourse, biodiversity is consistently undervalued, not least because of the conflict between economic development and habitat conservation, as Aldo Leopold noted in 1949. Capitalism and free-market enterprise are not best-fitted to place an appropriate value on what have been termed ecosystem services, such as the provision of oxygen to the air we breathe as a by-product of photosynthesis. It is relatively easy to understand the significance of, say, ozone depletion (hence the rapid international agreement to sign the Montreal Protocol phasing out ozone-depleting

Table 26.2. *Major soil groups in drylands and their susceptibility to various forms of land degradation. (Compiled by the author, based on his field observations in Africa, Asia and Australia.)*

Parent material	Soil group	Natural erosion status	Desertification hazards
Fresh or weathered rock	Cambisols	A1	A2, B1
	Leptosols	A1	A2
	Regosols	A1	A2
Weathered rock	Ferralsols	A1, B1	A2
Limestone	Calcisols	A1	A2
Sand sheets	Arenosols	C1	A1, B1, C3
Alluvium	Fluvisols	A1, B1	A2, B1, D3
	Luvissols	A1, B1	A2, B2, C2, D3
Clays	Vertisols	A1, B1	B2, D2
Sands over clays	Planosols	A1, B1	A2, B3, C1, D2
Clays, silts, sands	Solonchaks	A2, B1, D1	A2, B2, D3
	Solonetz	A2, B1	A3, B3, D3

A: sheet erosion; B: gully erosion; C: wind erosion; D: salinization

1: minor; 2: moderate; 3: severe

substances), but it is far more difficult to come to grips with the full complexity of biodiversity and its operation at the three different levels of ecosystems, species and genes. The 2005 Millennium Ecosystem Assessment painted a bleak picture of the global threats to species extinction from habitat destruction, the introduction of invasive species, pollution, the over-exploitation of biological resources and climate change. It follows that any actions that are taken to prevent or mitigate desertification processes will also reduce present-day threats to biodiversity in the desert world.

26.4 Conserving soil and water in drylands

Water, rather than soil, is almost invariably the limiting factor controlling land use in dry areas (Williams and Balling, 1996; Barakat and Hegazy, 1997). However, the nature of arid land soils can be extremely variable and needs to be taken into account when planning different forms of land use, whether irrigation, livestock rearing or rain-fed cultivation. Table 26.2 is a qualitative portrayal of the erosion hazards and susceptibility to different forms of land degradation of arid zone soils in Africa, Asia and Australia based on the author's field observations over many years. If we compare the natural erosion status listed for different soils in column three with the desertification hazards listed in column four, it is clear that these soils are often very



Figure 26.1. Reclaimed gully, middle Son Valley, north-central India.

vulnerable to accelerated soil erosion from wind and water and/or salt accumulation, depending on the type of soil.

Fortunately, even quite simple measures can be used to minimise soil erosion. In North Africa, the Romans built stone dams across the coastal stream channels to trap sediment behind the dams, which in turn trapped water within the sediments, so wheat could be grown and olive trees planted. Two thousand years later, the farmers in the arid limestone country of Tunisia use the same technique, and flourishing groves of date palms or olive trees may be seen in the valley bottoms. In the gullied lands adjacent to the Son Valley in semi-arid north-central India, the local peasant farmers build check dams of soil eroded from the gully walls and obtain two crops a year from the small fields within the gullies upstream of each earth dam (Figure 26.1).

Local tradition has it that the gullies only began to develop after clearing of the native vegetation during the British occupation, which seems plausible.

Another very effective method widely used on steep hill slopes in the drylands of South America, Africa, the Middle East and Asia is terracing. Provided the terraces are well-constructed and well-maintained, and some natural fertilizer is returned to the soils, the fields behind the terraces can be cultivated for centuries, or even millennia, as in parts of the Negev Desert highlands in southern Israel and along the flanks of Jebel Marra volcano in western Sudan. Some of the deepest and most spectacular gullies in the world are those entrenched into the highly erodible loess soils of central China. Here the central government co-opted local peasant labour for one month each year to construct new terraces and repair damaged ones – a laborious task but one acknowledged somewhat grudgingly by the local people as necessary.

Minimum tillage is another approach to preserving some form of surface cover throughout the year in order to avoid the exposure of bare soil to erosion from raindrop impact and run-off. Although the yields from cereal crops in wet years may not be as high as they are when the stubble is burnt and pesticides and herbicides are applied to eliminate weeds and pests, they remain relatively stable during both wet and dry years, because the soils retain a larger stock of water beneath the stubble mulch than they do when bare.

Central to all effective land use is the need for an effective exchange of information between farmers and agricultural extension officers. One initiative developed in Australia three decades ago and still active is the *Landcare* movement, which at its most successful allows new ideas and techniques to be widely disseminated among farming communities about how to improve agricultural yields while maintaining or enhancing soil fertility on a long-term basis (Campbell and Siepen, 1994). Another effective approach to achieving sustainable land use in arid areas is the provision of accurate and timely information about the likelihood of above- or below-average rainfall in order to allow farmers to plan their sowing and harvesting dates more strategically (Boyd et al., 2013). Given a suitable farming environment, crop modelling software packages are another useful way of developing probabilistic scenarios about the potential impact of possible future changes in climate and in climate variability on crop yields (Luo et al., 2007; Luo et al., 2010).

At a regional scale, it is often helpful to be able to predict the possible impact of future climate change by extrapolating from long-term trends in temperature and rainfall and by using a modified version of the type of risk magnitude and frequency matrix that is dear to the hearts of occupational health and safety officials and was long used in the insurance industry. The aim here is to assess vulnerability of particular natural resources to future change by considering exposure to risk, sensitivity, potential impact and adaptive capacity (Bardsley, 2006), using a qualitative scale from low to high, much as the one used in Table 26.2 in relation to soil erosion hazards and risks in arid lands.

26.5 Irrigation, dams, reservoirs and disease

Irrigation has been practised in south-west Asia for at least 7,000 years. The world's oldest recorded dam was built south of Memphis in Egypt under the aegis of Pharaoh Menes for water supply and irrigation. A ceremonial stone mace-head shows an Egyptian ruler excavating a canal sometime between 5,400 and 5,000 years ago (see Butzer, 1976, p. 21, fig. 2). Basin irrigation of wheat and barley was underway in the Nile Valley by about 6,500 years ago, and irrigated farming was practised in south-west Asia at least 1,000 years before then. The primary reasons behind the adoption of irrigated farming remain the same today as they were 5,000 years ago, namely, to cope with the vicissitudes of sparse and erratic rainfall and to capitalise on regular floods from rivers flowing through desert lands. However, what has changed today is the scale of dam building and the alternative uses to which the dammed water may be put, notably, to generate hydroelectric power. There are certain long-understood environmental costs associated with the construction of dams and reservoirs in the drier regions of the world (Kassas, 1972), of which the spread of water-borne disease is perhaps the most serious. The spread of Schistosomiasis (or Bilharzia) as a result of the reduction in river-flow velocity and the creation of artificial lakes has been a scourge in many parts of Asia, Africa and South America. The disease is caused by infestation with parasitic blood flukes transmitted by aquatic snails (Williams and Balling, 1996, pp. 137–140). An infected human can harbour about 4,000 pairs of flukes/worms, leading to destruction of the affected tissues. The parasites do not kill their human host but do cause weakness and anaemia from blood loss. Various chemical methods have been used to kill the snail carriers, or vectors, in the water, none of which are cheap, and some methods of biological control appear promising, in that they do not kill fish (Lemma, 1973). Other diseases associated with the presence of large bodies of standing water are malaria (the mosquito larvae live at the water surface) and filariasis, which is transmitted by a species of fly that lays its eggs in still water. All too often in their enthusiasm for dam building, government authorities fail to factor in the costs of disease and to make disease prevention an essential part of their planning.

26.6 Keeping fossil groundwater use in balance with aquifer recharge

We have seen in earlier chapters that the contrast between the late Pleistocene (LGM) aridity of the Sahara, the Gobi and the Thar deserts and their well-vegetated and well-watered early Holocene condition had a considerable impact on the Late Stone Age/Upper Palaeolithic and early Neolithic peoples who benefited from these changes. As the great continental ice sheets melted, postglacial temperatures and sea levels rose around the world. Evaporation from the intertropical oceans increased once sea surface temperatures became warmer. The summer monsoons of tropical Australia,

Asia and Africa brought reliable rainfall to the seasonally wet margins of the tropical deserts. Aquifers were replenished and groundwater levels rose, sometimes feeding springs and small lakes. Formerly dry lakes refilled. The mobile late Pleistocene dunes became vegetated and stable. Savanna woodland and grassland reoccupied what are today the semi-arid regions of the world. Rivers flowed again in many parts of the Sahara and Arabia. Countless small freshwater lakes supported communities of Upper Palaeolithic/Late Stone Age hunter-fisher-gatherer people who left behind their middens and hearths as enduring testimony to these brief episodes when the deserts were green once again.

Some groundwater recharge also took place during previous interglacial periods. However, because each interglacial episode was of relatively short duration, or roughly one-tenth of each glacial-interglacial cycle during the past million or so years, the duration of these recharge intervals was quite short. The consequence of all this is that the groundwater in major aquifers, like the Great Artesian Basin in central Australia or the Nubian Sandstone aquifers in North Africa, may be up to a million years old. Current use of the fossil groundwater in Australia, Arabia and in some of the North American deserts is well in excess of current recharge rates. In essence, we are now mining a non-renewable resource.

26.7 Importance of avoiding pollution of water, soil and air in drylands

In the case of groundwater, it is not just the quantity used that matters. Equally important is the actual quality of the water. Groundwater in the drier parts of northern India and northern China is often severely polluted by industrial effluents and by leakage from polluted rivers of toxic chemicals and untreated sewage. The same is true of many other dryland regions in the world where environmental controls and monitoring are inadequate or non-existent. The result is many hundreds of millions of deaths from poisoning and from gastro-intestinal illnesses.

The abundant use of detergents and of fertilisers containing phosphates and nitrates can lead to river pollution from cyanobacterial or algal blooms, particularly in turbid desert rivers, where the seasonal river discharge can be sluggish. Algal blooms are not confined to rivers but can be a hazard in coastal regions where the rivers carry dissolved loads of nitrates and phosphates.

Perhaps the most widespread form of water pollution in arid areas is from salt. We are only now starting to come to grips with the full complexity of how salt enters the groundwater system and how it is transported from groundwater to rivers and soils (Lawrie and Williams, 2004). As long as evaporation greatly exceeds precipitation, leaching of salt from the land surface will be difficult. It does not make sense to use groundwater to cultivate crops on a large scale in fully arid areas where the potential evaporation rates may amount to 5,000 mm/year or more, as at Kufra Oasis in south-east Libya, given that all water contains some dissolved salts, and these salts will

Table 26.3. *Major global environmental issues specified by Tolba and El-Kholy (1993)*

1. Air pollution
2. Ozone depletion
3. Climate change
4. Availability of freshwater
5. Coastal and marine degradation
6. Land degradation
7. Deforestation and habitat loss
8. Loss of biological diversity
9. Environmental hazards
10. Toxic chemicals and hazardous wastes

precipitate out on the surface and accumulate within the top-soil, eventually killing plant growth.

26.8 Importance of poverty alleviation, health and food security in drylands

It is interesting to note how widely perceptions vary in regard to global environmental issues. Table 26.3 provides a list of ten key issues regarded as significant by Dr Mostafa Tolba, distinguished microbiologist and for seventeen years the Executive Director of the United Nations Environment Programme (UNEP). The list is not in any order of priority but simply provides the topics discussed in each chapter of the book (Tolba and El-Kholy, 1993). Few would disagree that they are important, but as we saw earlier, there are more fundamental issues at stake, notably poverty alleviation, health and food security (Table 26.4). If government regulations and institutionalised corruption do not allow poor farmers to market their produce at fair and competitive prices and to retain and store the surplus, then they will remain trapped in poverty, as they are, for example, in many parts of Africa, South America and Asia. The result will be malnutrition, poor health and little incentive and ability to innovate and work more efficiently. Central to this is the widespread marginalisation of women and low levels of education and literacy, which lead to a vicious circle of poor hygiene, high levels of child death and the desire for large numbers of children to provide some form of safety net when the parents are no longer able to work. These factors are, of course, not peculiar to the drylands, but they are exacerbated by the remoteness of many desert regions and the difficulties of access. However, the situation is not entirely bleak. For example, in the Limpopo Basin of Botswana, with its highly variable climate, the local people have developed a body of indigenous knowledge, institutions and practices that enable them to cope successfully with natural hazards, such as major droughts (Dube and Sekhwela, 2008). Dube and Sekhwela (2008) also show how this knowledge base is in danger of being eroded, leaving the poorer communities more vulnerable than they previously were.

Table 26.4. *Primary aims decided at the Desert Margins Initiative, Nairobi, 23–26 January 1995. (Adapted from International Crop Research Institute for the Semi-Arid Tropics, 1995 and author's notes.)*

Overall goal

To contribute to sustainable food security and poverty alleviation

Overall objective

To promote innovative and action-oriented dryland management research to arrest land degradation

Specific objectives

Understanding land degradation

Assessing dryland management practices

Evaluating the role of livestock in the rangeland/arable land continuum

Designing policies, programs and institutional options

Improving natural resource management

Fostering domestication of tree species

Formulating drought management strategies

Enhancing institutional capacities

Exchanging technologies and information

26.9 Conclusion

Given the widespread problems of land degradation and desertification discussed in [Chapter 25](#), the question that comes most immediately to mind is whether or not we can indeed live in an ecologically sustainable manner in the deserts and their semi-arid and seasonally wet margins. Our survey of the past offers grounds for cautious optimism. Plants, humans and other animals have adapted in a variety of effective and often highly ingenious ways to the extremes of life in deserts, where water is, and always will be, the limiting factor governing life in these harsh environments. Faced in the past with extreme events, such as prolonged drought, all living organisms have responded in one of three ways. Some have migrated into less harsh environments, as did the nomadic Neolithic pastoralists who moved out of the Sahara when it dried out some 4,500 years ago. Others adapted and remained, as did the San hunter-gatherers of the Kalahari. Still others neither adapted nor migrated and instead became extinct.

Four conditions seem necessary if we are to continue to live in harmony with the desert world. First and most fundamental is the need to conserve desert ecosystems, since the only means of achieving a net increase in primary productivity on this earth is through photosynthesis, which requires plants to be able to capture the energy from the sun and carbon dioxide from the atmosphere to manufacture additional plant food in the form of starch. Second, taking due heed of the first law of thermodynamics, we should not systematically *remove* materials from any natural system or humanly modified system at a rate that is faster than the rate at which that system can replenish those materials. Extracting groundwater at a rate faster than its recharge rate disobeys this principle, as does allowing soils to erode at a rate faster than they can develop.

Third, we should not systematically *add* materials to any natural or humanly modified system at a rate faster than the rate at which that system can absorb and recycle those materials. Pollution of soil, water and air from pesticides, herbicides, fertilizers, chemical effluents and soot particles are all examples of infringing this principle. Finally, simple social justice decrees that all people should have fair and effective access to the natural resources of this earth in order to ensure better health, food security and poverty alleviation. Reliable and easy access to clean and safe drinking water is a case in point. The quotation at the head of this chapter implies that unless reverence for all forms of life underpins our approach to natural resource management, economics alone can never achieve ecologically sustainable use of our deserts and their margins. White (1967) made the same point nearly half a century ago.

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Index

- Accelerated soil erosion, 70, 166
- Acheulian, 118–19, 198, 203, 271, 308–9, 313, 345
- Adaptations to aridity, 37–56
 - animals, 45–6
 - humans, 37–40, 203, 216–55
 - plants, 40–6, 62
- Adiabatic lapse rate, 14, 234
- Adrar Bous, Sahara, 102, 189, 271, 312, 317–18
- Aeolian/eolian dust, *see* Dust
- Afar desert, *see also* Middle Awash
 - alluvium, 159–60
 - hominids, 159, 286–7, 304
 - lakes, 65, 159, 190–1
 - rift, 337–9
 - stone tools, 32
- Africa, *see also* Sahara
 - biomass burning, 484–6
 - Cenozoic desiccation, 30, 331–7
 - Cenozoic tectonism, 30, 331–7
 - Chad basin, 30–1
 - climate, 329–30, 470–1
 - desertification, 317, 494–6
 - drought, 464–8
 - dust, 149
 - glaciation, 225–6, 235–6
 - Harmattan, 72, 142
 - Holocene climates, 184–5, 189–90, 222–3, 470–1
 - lake fluctuations, 30–1, 191, 198–202, 215–18
 - last glacial maximum, 220–2
 - overgrazing, 317, 488–9
 - precipitation, 329–30
 - Quaternary climates, 215–18, 224, 344
 - tectonic history, 25, 30, 331–7
 - volcanoes, 339–40
 - wind systems, 329–30
- Afro-Arabian dome, 337–3
- Agriculture
 - origins, 47, 311–13
 - prehistoric, 47, 50, 311–13
- Aïr Mts, 30
- Albedo, drought model, 466–8
- Allerød-Bølling interstadial, *see* Bølling-Allerød interstadial
- Alluvial fans, 169–70, 172, 178, 266–7, 461
- Alpine glaciations, 215–16, 226–7
- Amazon basin, 75, 408–9, 415–17
- America, *see* North America, South America
- Amino-acid racemisation dating, 95–6
- Andes, uplift, 239–40, 407–13
- Antarctica
 - carbon dioxide, 35, 243, 505
 - cooling, 28
 - dust, 29
 - glaciation, 24, 28, 30, 243–4, 431
 - ice core record, 35, 228, 505–6, 525
 - methane, 35, 505
- Aphelion, *see* astronomical theory
- Aquatic organisms, terrestrial, 289–93
- Aquifers, recharge, 40
 - see also* groundwater
- Arabian deserts, 67, 349–51
- Aral Sea, 490–1
- Archaeological evidence, 308–27
- Arctic sea ice
 - present extent, 513
 - Cenozoic growth, 391, 393–4
- Ardipithecus*, 305
- Areic, 38, 162
- Argon-argon dating, 85–6
- Aridity
 - causes of, 14–18, 20–1
 - glacial, 185–6, 220–2
 - present-day distribution, 2
- Arrhenius, Svante, 508–9
- Arroyo, *see* Gullies
- Asia, *see also* China, India
 - Cenozoic cooling, 366–7
 - Cenozoic desiccation, 32, 364–8
 - Cenozoic tectonism, 364–8
 - climate, 361–3
 - deserts, 68, 360–83
 - desertification, 68, 489–94
 - dunes, 370–2, 491–4
 - glaciation, 237–9, 378–9
 - lakes, 204–5, 218, 377–8
 - landscapes, 363–4

- Asia (*cont.*)
 loess, 491–2
 marine sediments, 381–2
 monsoon systems, 361–3, 367–8
 onset of desiccation, 364–8
 rivers, 186, 218–19, 366, 372–5
 speleothems, 380–1
 Toba volcano, 375–7
- Astronomical cycles, *see* astronomical theory
- Astronomical theory (of climatic change), 33, 227–8
- Atacama desert, 17, 75, 408–13
- Atlantic Ocean
 influence on climate, 229, 240, 244, 391
 thermohaline circulation, 391
- Atlas Mts., glaciation, 225
- Atmosphere
 effect of composition on climate, 500, 503–10
 human impact on, 503–10
- Atmospheric carbon dioxide, 500, 503–10
- Atmospheric circulation, 14–16, 358, 410, 426, 451
- Australia
 biomass burning, 444–6, 485
 Cenozoic desiccation, 32, 298, 430–4
 Cenozoic tectonism, 25–7, 430–3
 climate, 423–7
 Cooper Creek, 72, 140, 159, 207, 440
 deforestation, 486–8
 desert dunes, 71, 136–40, 436
 desert dust (loess), 70–2, 143, 154–5, 436–7
 desert exploration, 70–1
 desert landscapes, 6–7, 424–5, 428–30, 437
 droughts, 451–60
 early Holocene, 445
 Eastern Highlands, 192, 208, 239
 El Niño, 452
 erosion rates, 430
 faunal extinctions, 207, 250–1, 320–3, 442, 444–7, 485
 fire, 322–3, 444–6, 485
 fire-stick farming, 485
 Flinders Ranges, 195, 207, 428, 435, 439, 447
 floods, 451–60
 glaciation, 176, 242–3, 438
 Great Artesian Basin, 40
 Lake Eyre, *see* lakes
 Lake Mungo, *see* lakes
 lakes, 72, 205, 207, 220, 435–6, 441–3
 last glacial maximum, 72, 206–7, 438–9, 441, 443–4, 448
 Lynch's Crater, Queensland, pollen record, 434–5
 Murray-Darling, 266
 periglacial solifluction, 72, 220, 438
 present-day climate, 423–7
 Quaternary climates, 219–20, 434–48
 Riverine Plain, 72, 174–6, 439–40
 rivers, 72, 174–7, 186, 439–41
 salinization, 488
 separation from Antarctica, 26–7, 431–2
 Snowy Mts., 176, 438
 source-bordering dunes, 140, 439–41
 summer monsoon, 447
 vegetation history, 186, 433–6, 438–9
 wetlands, 437
 westerlies, 437
 Willandra lakes, 192–3, 205–6, 208, 437, 443, 437
Australopithecus, 53, 159, 286, 300
- Badain Jaran desert, China, 204, 364, 371–2
- Bale Mts., Ethiopia, glaciation, *see* Ethiopia, glaciation
- Beryllium 10 (¹⁰Be) dating, 94–5, 237–8, 240–2, 305, 366, 428–30
- Bilharzia, *see* schistosomiasis
- Biodiversity, 519–23, 528
- Biogeophysical model of drought, 466–8
- Biomass burning, 484–6
- Bioturbation, *see* soil, bioturbation
- Bipolar seesaw, 241, 243–4
- Blue Nile
 alluvial history, 105, 181–4, 186, 217, 267, 342, 351–4
 distributary channels, 171–81
 fan, 177–81
 gorge, 342–3
 incision, 178, 342–3
- Bolivian Altiplano, 34, 408, 412–14
- Bølling-Allerød interstadial, 256, 297, 324, 405
- Bond cycles, 36
- Botswana, 468, 522
- Brazil
 droughts, 468–9
 speleothems, 255–6, 468–9
- Brunhes magnetic chron, 83–4
- Butchery site, prehistoric, 318–20
- Cainozoic, *see* Cenozoic
- Calcrete, 160, 219, 275–88, 370
- Carbon dioxide (CO₂)
 glacial atmospheric levels, 506
 historic increase in atmospheric concentration, 505
 interglacial atmospheric levels, 506
 possible impact on future global temperature, 505
 released by burning fossil fuels, 507
- Carbon isotopes
 in plants, 31, 107, 367
 in soils, 31, 107, 109, 367–8, 370, 375, 395
 in speleothems, 247, 251–3, 256
- Cenozoic
 cooling, 28
 desiccation, 24, 29, 30
 sedimentation, 27
 tectonism, 24–30
 time-scale, 23
 weathering, 280, 298
- Charophyte, 390
- Chihuahuan desert, *see* North America
- China
 Cenozoic cooling, 28
 Cenozoic desiccation, 24, 29–30
 desert dunes, 131–4, 370–2
 desertification, 491–4
 droughts, 452–9
 floods, 452–9
 lakes, 204, 377–8
 loess, 31–2, 35, 69, 152–4, 157, 368–70

- Loess Plateau, 32, 152–4, 157, 366, 368
- speleothems, 251–3, 380–1
- Chironomid, 290–1
- Cladocera, 290
- Clay dunes, 121–2
- Climate change, present-day
 - causes of, 500–15
 - carbon dioxide, 500, 504–7
 - greenhouse effect, 500, 503–10
 - impact, 500–1, 510–15
 - methane, 504–7
 - trends, 505, 508–10
- Climatic cooling and volcanism, 517
- Climatic warming, *see* climate change, trends
- Clumped isotopes, 99
- Colorado plateau, *see* North America
- Complex response, of rivers, 9
- Continental breakup, 25–6, 431–2
- Continentality, 16–17
- Cores, deep-sea, 381, 501
- Cosmogenic nuclide dating, 94–5
- Cosmogenic radionuclides (CRN), 94–5
- Dams, 526
- Dansgaard-Oeschger (D-O) events, 35, 248, 251
- Darwin, Charles, 70, 75, 142, 146, 407
- Dating methods, 77–97
- Dead Sea, Lake Lisan, 197, 202–3
- Dead Sea rift, 273
- Debris-flow, 169
- Deep-sea cores, *see* cores, deep-sea
- Deep weathering, *see* weathering
- Deflation, 191
- Deforestation, 486–96
- Deglaciation, causes of, 228–9
- Dendrochronology, 96
- Denisovan, 307
- Desert
 - causes, 14
 - definition, 3, 12
 - distribution, 2, 13
 - dunes, 112–41, 370–2
 - dust, 142–57, 363, 366, 368–70
 - encroachment, 475
 - floods, 158–9
 - glaciation, 224–44
 - history of research, 57–76
 - lakes, 8–9, 189–223
 - landscapes, 1, 4
 - origins, 516
 - pavement, 167, 273
 - pluvial, 209–23
 - polygenic landscapes, 4, 5–8
 - refugia, 284–5
 - rivers, 9, 38–9, 158–88
 - soils, 258–82
- Desertification
 - Africa, 494–6
 - Asia, 489–94
 - assessment, 496–7
 - Australia, 486–9
 - biomass burning, 484–6, 499
 - biophysical factors, 485
 - causes, 66, 477, 480–6, 517, 519
 - consequences, 477, 484, 490
 - definition, 473–6, 486, 517
 - detection, 488
 - dryland salinization, 487–8
 - economic factors, 481–3
 - extent, 489, 496–7
 - fire, 484–6, 495
 - hazards, 523
 - human impact, 473–99
 - natural, 476, 477–9
 - overgrazing, 488–9, 493
 - plant cover, 479–80, 487–8, 493–6, 520–3
 - prevention, 497–9, 516–30
 - processes, 476–9, 480–6
 - social factors, 481–3
 - solutions, 498, 516–30
- Deuterium, temperature record, 243
- Dew point, 15, 16
- Diatoms, 65, 142, 146, 189, 191, 194, 299–300, 345
- Disjunct distribution, fauna and flora, 284–5
- Distributary/distributive fluvial system, 162, 164, 173
- Drake Passage, opening of, 24, 367, 412
- Droughts
 - albedo model, 466–8
 - causes of, 449–72
 - definition, 450
 - El Niño–Southern Oscillation (ENSO) events, 451–62, 471–2, 511–12
 - historic, 60, 158, 449–72
 - Sahel, 464–8
 - volcanic eruptions, 462–4
- Dunes
 - Active, 19, 114–15
 - Africa, 119–20, 125, 134
 - Arabia, 126–7
 - Australia, 19, 136–40
 - barchan, 113, 118–19
 - China, 131–4
 - clay, 121
 - dating, 119, 121–3
 - distribution, 114–17
 - downwind of lakes, 137
 - fixed, 114–16, 119, 125, 266
 - gypsum, 121
 - Kalahari desert, 134–6
 - linear, 113, 117–19, 137
 - lunette, 121–2
 - mobilisation, 20, 117–19, 492–4
 - Namib desert, 134–6
 - Negev desert, 126
 - North America, 141
 - orientation, 119
 - soil, 265–6
 - stabilisation, 125–6
 - Sahara, 19, 30, 119, 123–6
 - source-bordering, 120–1, 140, 352–3
 - South America, 409, 417
 - Thar Desert, India, 127–30
- Duricrust, 258, 275–80

- Dust
 - characteristics, 143, 145, 147
 - climate, effect on, 154–6
 - deposition, 146–7, 149, 151, 154, 273
 - distribution, 144–5, 150, 152
 - entrainment, 147–8, 274
 - in ice cores, 145, 156
 - in marine cores, 145, 148, 221
 - in soils, 265–6, 273–4, 278
 - origins, 145
 - particle size, 143, 146
 - plumes, 144–5
 - rainfall, relation to, 149
 - reworking, 146, 151
 - sources, 146, 148–9, 154–5
 - transport, 147, 149
 - see also* Desert dust
- Dust Bowl, 512
- Dust, volcanic, effect on climate, 462–4
- East Africa, Cenozoic desiccation, 340–1
- Ecological sustainability, prerequisites, 516–30
- Effective precipitation, 4
- El Niño–Southern Oscillation (ENSO) events, 317, 451–60, 471–2, 485, 511
- Electron spin resonance (ESR) dating, 94
- Eluviation, *see* soil
- Endoreic, 38–9, 162
- Eocene–Oligocene transition, 366–7
- Epic of Gilgamesh, 57, 209
- Equifinality, 9, 270
- Equilibrium line altitude (ELA), 234, 240–1, 244
- Erosion
 - accelerated, 166
 - geological, 166
 - processes, 166
 - surface, 167
- Ethiopia
 - Afar desert, 190, 338–9
 - deforestation, 494–6
 - glaciation, 225–6, 235
 - lake fluctuations, 185, 190–1, 339
 - Rift Valley, 339–40
 - volcanoes, 339–40
 - uplift, 341–3
- European Alps, 215–16, 226–7
- Evaporation, 3, 207, 470
- Exoreic, 38, 162
- Extinctions
 - Australia, 320–3
 - Blitzkrieg model, 323–4
 - causes, 320–4
 - Eurasia, 324
 - North America, 323–4
 - prehistoric fauna, 320
- Ferricrete, 275–6, 278–80
 - see also* laterite
- Fire
 - biomass burning, 484–6
 - prehistoric, 55, 484–5
- Floods
 - causes of, 449–72
- El Niño–Southern Oscillation (ENSO) events, 451–60, 471–2
 - flash, 449–50
 - historic, 158–9, 450–72
- Flowstone, 246, *see also* speleothems
- Foggara, 39, 552–3
- Food security, 528–30
- Foraminifera (forams)
 - assemblages, 288–9, 298
 - isotopic composition, 227
 - oxygen isotopes, 227, 289
- Fossils, 283–301
 - invertebrates, 189, 194, 196, 267, 288–93
 - plants, 271–2, 295–300, 344–5
 - pollen, 195, 285, 294–9
 - vertebrates, 189, 283, 284–8, 307, 312
- Fossil fuel burning, 503–10
- Full glacial, *see* glacial maximum
- Gastropod, 178, 183, 195–6, 267, 292–3
- Geomagnetic dating, 83–4
- Gezira plain, Sudan, 158, 177–8, 186, 292–3
- Gibber, 18, 436
- Glacial
 - aridity, 185–6, 206, 220–2, 206
 - cycle, sawtooth pattern, 34
 - erosion, 226, 229–32
 - erratic, 231–2
 - maximum, 238, 398
 - moraine, 230–3, 237
 - pluvial, *see* pluvial
 - outwash, 232
 - stages, North America, 396–8
 - striations, 230–1
 - termination, 229, 237
 - theory, *see* astronomical theory
- Glacial-interglacial cycles, 27, 34, 226–9
- Glaciation
 - Africa, 235–6
 - Asia, 237–9
 - Australia, 242–3
 - causes of, 33, 227–8
 - evidence of, 229–32, 236
 - chronology, 226–9
 - Europe, 396
 - North America, 239, 241–2
 - South America, 239–41
- Global Assessment of Soil Degradation (GLASOD)
 - study, 496–7
- Global climates, present, 2, 15, 16
- Global conveyor belt, *see* thermohaline circulation
- Global cooling, 29, 517
- Global ice volume and oxygen isotopes, 101
- Global warming, 508–10
- Gobi, 18
- Gobi desert, 364
- Goyder's Line, 41–2
- Grand Canyon, *see* North America
- Great Artesian Basin, Australia, 40
- Great Basin, USA, *see* North America
- Greenhouse effect, 243, 500, 503–10
- Greenhouse warming, 508–10
- Greenland ice core records, 35, 225, 228, 256

- Groundwater, 40, 42, 526–8
- Gullies
- alluviation, 169–74, 401–2
 - incision, 168, 169–74, 401–2, 478–9
 - vegetation, impact of, 171–4, 401–2
- Gypcrete, 275–6, 281
- Hadley circulation, 14–15, 410, 451
- Half-life, 80
- Hamada, 18
- Harappa, 375
- Harmattan, 72, 144
- Heinrich events, 35–6, 222, 248, 256, 358
- Herodotus, 1, 58, 159, 177, 351
- Hillslope erosion, 366–7
- Himalayan uplift, 25, 30
- Hoggar Mts., Sahara, 5, 30
- Holocene, subdivision, 22
- Hominid, 302–7
- Hominoid, 288
- Homo erectus*, 203, 305, 313–4, 326
- Homo ergaster*, 306, 313–14, 326
- Homo habilis*, 305–6
- Homo heidelbergensis*, 307
- Homo neanderthalensis*, 307
- Homo sapiens*, 307, 314–6, 351
- Human health, 51, 526, 529–30
- Hydrogen isotopes, 243
- Hydrological cycle, 29
- Ice cores
- Africa, 225
 - Andes, 479
 - Antarctica, 505, 507
 - Greenland, 380
- Ice-rafting, 27, 35
- Ice sheet
- Late Pliocene build-up in Northern Hemisphere, 24, 27, 30
- Ice volume, global, 101
- Illuviation, *see* soil
- India
- calcrete, 109, 130
 - convergence on Asia, 24
 - desert dunes, 109, 127–30, 370
 - desert lakes, 218–19
 - droughts, 452–9
 - rivers, 186, 372–5
 - separation from Australia, 24–5
 - summer monsoon, 452–9
 - Toba volcanic ash, *see* Toba volcano
- Indian Ocean, 454, 460–1
- Indian Ocean Dipole (IOD), 454, 460–1
- Indus Valley Culture, 69, 218–19, 375
- Infiltration, 165, 168
- Inselberg, 48, 167–8, 259
- Interdecadal Pacific Oscillation (IPO), 469
- Interglacial, 34, 224–5, 353
- Interstadial, 35, 297
- Intertropical convergence zone (ITCZ), 15, 16, 185, 358, 410, 426
- Irrigation, 50–1, 526
- Invertebrates, 267, 288–93
- Isotopes
- carbon, 53, 102–4, 106–9, 196, 247–56, 293
 - clumped, 99, 249
 - deuterium, 243, 247
 - marine stages, 84–5, 227
 - oxygen, 100–4, 107, 196, 247–56, 293
 - marine stratigraphy, 84–5, 100–1
 - strontium, 99, 178, 187, 376
- Isotopic equilibrium, 80–1
- Isotopic fractionation, 85, 100, 106–7
- Israel
- Desert dunes, 350–1
 - Negev desert, 350–1
 - Pleistocene Lake Lisan, 197, 202–3
 - speleothems, 350–1
- Jebel Marra volcano, Sudan, 30, 270–1, 344
- Jet stream, 30
- Jet stream, easterly, 30
- Kalahari desert, 32, 55, 134–6, 354–9
- Kenya, lake fluctuations, 185–6, 196, 199, 215–18, 346–7, 465
- Kilimanjaro, glaciation, 225, 233, 236
- Lake
- amplifier, 193
 - classification, 193
 - fluctuations, 194, 197–207
 - organisms, 191, 194–5, 197
 - origins, 190
 - regression, 191, 194
 - reservoir, 193
 - saline, 194, 367, 435
 - salinity, 194, 527–8
 - sediments, 191, 194
 - shorelines, 192
 - stable isotopes, 196–7
 - temperature, 194, 197
 - transgression, 191, 194
 - water budget, 192–3
- Lakes
- Abhe (Afar desert), 65, 159, 254–5
 - Bonneville (USA) glacial lake, 74, 211–13, 401–3
 - Chad (Africa), 64, 160, 194, 200, 207, 222, 465
 - China, 68, 203–5, 377–8
 - Didwana (India), 219
 - Eyre (Australia), 72, 159, 194, 205–7, 442–3
 - Frome (Australia), 139–40, 207, 436, 446–7
 - Gadeb (Ethiopia), 30, 191, 299
 - George (Australia), 72, 230
 - Lahontan (USA) glacial lake, 211, 213–14, 402–4
 - Lisan (Israel), 197, 202–3
 - Makgadikgadi (Africa), 200–1, 208
 - Malawi (Africa), 201, 359
 - Masoko (Africa), 201–2, 359
 - Mungo (Australia), 155, 205
 - Ngami (Africa), 194
 - Shati (Africa), 198, 346
 - Tana (Africa), 200, 221
 - Tibetan plateau, 204–5, 377–8
 - Victoria (Africa), 200, 223

- Lakes (*cont.*)
 White Nile, last interglacial lake, 195
 Zihor (Israel), 203
 La Niña, 159, 452
 Last glacial maximum (LGM), 34, 152, 195–9, 205–6, 213, 217, 221–8, 238–9, 285, 289, 350, 358, 416
 dust flux, 148, 205, 222
 sea level, 443
 Late Cenozoic
 cooling, 24, 29
 desiccation, 24, 29
 ice build-up, 24
 vegetation history, 24, 31
 Laterite, 69–70, 264, 275–6, 278–82, *see also* ferricrete
 Laurentide ice cap/sheet, *see* North America
 Levallois, 309–10
 Lithospheric plate movements, 25–6
 Little Ice Age, 464
 Locusts, 50–1
 Loess, *see* Dust
 Luminescence dating, 92–4, 198
 Lunettes, 121–2
- Macrofossils, plant, 293–5
 Magnetopolarity time scale, 83–4
 Makgadikgadi basin (Africa), *see* lakes
 Marine isotope stage (MIS), 84–5, 227, 248
 Matuyama magnetic chron, 83–4
 Meat-eating, prehistoric, 318–20
 Mediterranean
 Miocene desiccation, 24, 333, 340
 sapropel record, 184–5, 351–4, *see also* Nile
 Megafan, *see* rivers
 Mesa, 39
 Mesolithic, 189, 311, 318
 Mesozoic, 430–2
 Messinian salinity crisis, 24, 333, 340
 Methane
 Atmospheric, 504
 ice core records, 506
 recent increase in atmospheric, 506
 Microfossils
 marine, 288–9
 non-marine, 295–300
 Microlith, 309, 311
 Middle Awash valley, 159–60, 191–2, 272–3, 287, 300, 305–8
 Migrations, prehistoric, 313–18
 Milankovitch cycles, *see* astronomical theory
 Minevaporal hypothesis, *see* pluvial
 Mitochondrial DNA (mtDNA), 313, 315–16, 327
 Mohenjo-Daro, 375
 Mojave desert, *see* North America
 Molecular biology, 285, 302, 313–16
 pongid and hominid divergence, 304
 Mollusca, non-marine, 291–3
 Monsoon
 African, 178, 196, 200, 329–30
 Australian, 427, 435, 447
 Asian, 204, 251–2, 361–3
 Moraine, glacial, *see* glacial, moraine
 Mount Kenya, glaciation, 225, 233, 236
 Mousterian, 309–10
 Namib desert, 17, 134–6, 159, 354–9
 Neanderthal, 307
 Negev desert, 146–7, 249–50, 273, 281, 350–1
 Neodymium isotope ratios, 265
 Neolithic, 47–8, 56, 66–7, 183, 189, 311–13, 317–19
 Nile, *see also* Blue Nile, White Nile
 alluvium, 351–4
 delta, 105–6, 185
 floods, 159, 351–4, 452–9, 462
 gorge (Egypt), 340
 Herodotus, 58, 159, 351
 response to vegetation changes in headwaters, 181–4
 sapropel record, Mediterranean, 184–5, 351–4
 Nitrogen isotopes in bones, 107
 Nomads, subsistence strategies, 47–50
 Normalised Difference Vegetation Index (NDVI), 479
 North America
 alluvial fans, 399–402
 arroyos, 168–74, 401–2
 Baja California, 389
 Basin and Range province, 388, 391, 399
 Cenozoic cooling, 395–6
 Cenozoic desiccation, 394–6
 Cenozoic tectonism, 391–3
 Chihuahuan desert, 389, 399, 405
 climate, 389–91
 Clovis hunters, 384, 404–6
 Colorado Plateau, 171, 386, 388–9, 391–3
 deserts, 384–406
 desert landscapes, 6, 385–9
 faunal extinctions, 74–5, 385, 404–6
 glaciations, 30, 239, 241, 396–8
 Grand Canyon, 73, 253, 385–6, 392–3
 Great Basin, 76, 213–14, 387, 389, 403
 Great Plains, 388, 391, 395, 399
 Great Salt Lake, 402
 lakes, 74, 211–14, 400–4
 Last Glacial Maximum climate, 397–8, 405–6
 late Quaternary vegetation history, 405
 Laurentide ice cap, 384, 397, 399
 loess, 398–9
 meltwater, 391
 Mojave desert, 380, 389, 399
 Nebraska Sand Hills, 399–400
 packrat middens, 405
 paleosols, 395
 pluvial lakes, 74, 402–4
 Quaternary climates, 396–405
 rivers, 393, 399–400
 Rocky Mountains/Cordillera, 239–40, 387–8, 391–4
 Sierra Nevada, 387–8, 391
 Sonoran desert, 389, 399, 402, 461
 speleothems, 253, 392, 405
 tectonism, 391–3
 vegetation history, 395, 405
 North Atlantic, 240, 244, 391
 North Atlantic Oscillation, 468–9
 Nubian Sandstone aquifer, 40
- Ocean currents, 17
 Oceans
 microplankton, 100, 288–9, 298

- Ogallala Aquifer, 40
 Okavango (Africa), 200–1, 206, 208
 Oldowan, 271, 308–9
 Optically stimulated luminescence (OSL) dating, 92–4
 Orbital eccentricity, *see* [astronomical theory](#)
 Ostracods, non-marine, 195, 289–90
 Outwash, glacial, 232
 Overland flow, *see* [runoff](#)
 Oxygen isotopes, *see* [isotopes, oxygen](#)
- Pacific Decadal Oscillation (PDO), 469
 Pacific Ocean
 sea surface temperature anomalies, 469
 Southern Oscillation Index, effect of, 451, 511
 Pacific Oscillation, 469
 Packrat middens, 405
 Paleocene-Eocene thermal maximum, 395
 Palaeolithic
 Lower, 310, 313
 Middle, 198–9, 272–3, 287, 310, 314, 326, 351
 Upper, 310–11, 314, 316, 324
 Paleochannel, *see* [rivers](#)
 Paleoclimatic record, evidence used to reconstruct, 196–7
 Paleodiet, isotopic evidence, 106–8
 Paleoenvironmental indicators, 196–7
 Paleomagnetism, 83–4, 195
 Paleosol, *see* [soil](#)
 Paleothermometer, ostracods, 110–11, 289–90
 Palmer Drought Severity Index (PDSI), 469–70, 512
 Palynology, *see* [pollen analysis](#)
 Panama isthmus, closure, 24, 412
 Pangea, fragmentation, 24, 412
 Paratethys Sea, 31
 Pastoralists, nomadic
 drylands, influence on, 488–9, 493
 subsistence strategies, 47–50, 521–3
 Patagonia
 dust, 155
 glaciations, 241, 244
 Patterned ground, 233
 Periglacial
 landforms, 176, 232–3, 235
 solifluction, 233, 235
 Permafrost, Pleistocene, 232, 384
 Peru, impact of El Niño-Southern Oscillation (ENSO)
 events, 159, 452
 Petra, Jordan, 50, 52, 67
 Phreatic zone, 245
 Photosynthesis, pathways, 31, 107, 109, 367–8, 370, 395
 Phylogenetics, 285
 Phylogeography, 285
 Phytoliths, 300
 Plant macrofossils, 293–5
 Plants, soil conservation, 487, 521–5
 Plate boundaries, 25, 431
 see also [lithospheric plate movements](#)
 Pleistocene, timescale, 22–3, 32
 Pluvial
 Asia, 218–19
 Australia, 72–3, 219–20
 Definition, 67, 210–11
 East Africa, 215–18
 Dead Sea, 211
 glacial, 185, 211–14
 India, 218–19
 North America, 74, 211–14
 South America, 214–15
 Pollen, 31, 195, 433
 Pollen analysis,
 interpretation, 285, 295–8
 methodology, 295–6, 298–9
 Potential evaporation, 469–70, 512
 Potassium-argon dating, 85–6
 Potwar Plateau (Pakistan), paleosols, 30, 367–8
 Poverty alleviation, 528–30
 Precessional cycle, *see* [astronomical theory](#)
 Prehistoric
 butchery, 318–20
 extinctions, 320–4, 326–7
 fire, 320–3
 hominids, 302–7
 migrations, 313–18
 stone tools, 32, 304–11, 324–6
- Quaternary
 climatic variability, 287, 518–19
 duration, 23
 environments, sources of data used to reconstruct, 10, 217–18, 300–1
 sea-level fluctuations, 518–19
 time scale, 2, 22–4, 32
- Radiocarbon (^{14}C) dating, 80, 86–91, 277–8, 316
 Raindrop impact, 167–8
 Rain-shadow effect, 17–18, 361, 410, 412
 Rajasthan desert, *see* [Thar desert](#)
 Red Sea, crossing, 313–15
 Red Sea Hills, 292
 Refugia, 284–5
 Regression, lake, 191, 194
 Relative humidity, 15
 Rivers
 anabranching, 172–3
 anastomosing, 173
 bed load/traction load, 173, 176, 180–3, 186
 braided, 177
 channel pattern, 162, 177, 181
 dissolved load, 166, 279
 distributary, 173
 ephemeral, 162–3, 169–74
 floodout zone, 164
 incision, 178, 185, 372–5
 meandering, 177, 179–80
 megafan, 164, 266–7
 metamorphosis, 72, 173–4
 mixed load, 176
 paleochannel, 72, 160, 174–81
 plant cover, 181–4, 186
 stream power, 171, 173, 212
 suspension load, 162, 166, 173, 176
 terrace, 161–2, 164, 187, 372–5
 traction load/bed load, 173
- Rub' al Khali Desert, 127
 Runoff, 165, 168–9, 197

- Sahara desert
 Cenozoic desiccation, 30–1, 331–7
 dunes, 186
 dust, 144
 former drainage network, 19, 63, 160–1, 315, 333–5, 351
 fossil fauna, 189
 glaciation, 225
 Holocene lakes, 63–5, 160, 184, 189, 222, 347–9
 onset of aridity, 31, 331–7
 Pleistocene lakes, 64, 198–200, 222–3, 345–9
 pollen analysis, 31, 65–6
 prehistoric human occupation, 1, 62, 189, 316–17, 345–9
 Quaternary glacial aridity, 185–6, 347–8
 rock art, 1, 18, 283–4
 topography, 332
 vegetation history, 31, 65–6, 335
- Sahel drought, 464–8
- Salinisation, 488–91
- Salt lake, *see* lake, saline
- Sand dunes, *see* dunes
- Sangamon interglacial, 396
- Sapropel, 184–5, 351–4
- Schistosomiasis, 51, 526
- Sea-ice, *see* Arctic
- Sea-level
 and ice volume, 229, 384
 at glacial maximum, 384
 fluctuations, 229, 384, 508
 last interglacial, 224–5, 508
- Semien Mts, Ethiopia
 glaciation, 216
 periglacial deposits, 216
- Serir, 18
- Sierra Nevada, 387–8, 311
- Silcrete, 275–6, 280
- Silk Road, 69
- Simpson desert, Australia, 137–9
- Slopewash, 168, 430
- Snowline, *see* Equilibrium Line Altitude (ELA)
- Soil
 bioturbation, 260, 264, 268–70, 523–5
 buried soil, 259, 271, 274, 281
 carbonate, 108–9, 261–2, 265–6, 272, 274, 277, 375
 catena, 260
 characteristics, 258, 261–2, 266–8
 classification, 261–3
 climate, 260
 climatic zonalism, 263
 conservation, 523–5
 crust, 167
 definition, 259–60
 degradation, 165
 desert, 258–82
 dryland, 258–82
 eluviation, 168
 erodibility, 167
 erosion, 166, 266, 430
 factors, 259–60
 fauna, role of, 260, 268–70, 523–5
 formation, 259–60, 263–5
 fossil, 54, 258, 270–5
 horizon, 260–1, 264
 illuviation, 168
 infiltration, 165, 167–8, 468
 mapping, 258
 micromorphology, 264–5
 moisture, 468
 paleosol, 54, 258, 270–5, 281, 368–9
 permeability, 168, 267
 polygenic, 264, 274
 processes, 263–5
 reg, 273
 salinization, 487–8, 519
 stone layer, 268–70
 termite activity, 264, 268–70
 toposequence, 259–60
 vertisol (cracking clay), 178, 261–2, 267–8
- Soil creep, 166, 169
- Solar radiation
 energy balance, role in, 502–3
- Solifluction, periglacial, *see* periglacial solifluction
- Sonoran desert, *see* North America
- South America
 Amazon basin, 75, 285, 408–9
 Andes, 239–40, 407–13
 Atacama desert, 17, 75, 408–13
 Bolivian Altiplano, 34, 408, 412–14
 Cenozoic cooling, 412–15
 Cenozoic desiccation, 412–15
 Cenozoic tectonism, 407, 412–15
 climate, 409–12
 desert exploration, 75
 dust/loess, 416
 El Niño, 407, 409, 451–9
 fossil soils, 414
 glaciations, 239–41, 417–19
 lakes, 214–15, 419
 landscapes, 407
 loess, 416
 Malvinas/Falklands current, 410
 Quaternary climates, 415–22
 paleosols, 414
 Patagonia, 75, 408–9, 412
 Peruvian/Humboldt current, 410
 pluvial lakes, 419, 421
 Quaternary climates, 415–22
quebrados, 414
 relict drainage, 414
 rivers, 408, 413, 420
salars, 408, 419
 vegetation history, 285, 415–16
- Southern Ocean, 24
- Southern Oscillation, 451
see also El Niño–Southern Oscillation
- Southern Oscillation Index (SOI), 451–60, 511
- Speleothems, 245–57
 Africa, 256
 Arabian peninsula, 248, 349–50
 Australia, 250, 435–6
 Brazil, 468–9
 China, 251–3
 definition, 245
 isotopes, 247, 251–3, 256
 Negev desert, 249–50, 350

- North America, 253
 - precipitation, 103, 246–7, 251–3
 - temperature, 103, 247
- Sponge spicule pottery, 60–1, 178
- Stadial, 35
 - see also* Younger Dryas
- Stick-nest rat midden, 435
- Stone Age
 - Early (ESA), 310
 - Middle (MSA), 198–9, 310, 314, 326
 - Late (LSA), 310–11, 326
- Stone tools
 - origins in Africa, 308–9
 - prehistoric, 32, 304–6, 308–11, 324–6
 - see also* prehistory, stone tools
- Stream power, *see* river
- Striation, *see* glacial striation
- Strontium/calcium ratios, 110, 290
 - see also* paleothermometer
- Strontium isotope ratios, 104–6, 376, 278, 265
- Strzelecki desert, Australia, 137–9
- Surface exposure dating, *see* cosmogenic nuclide dating
- Sustainability, 516–30

- Taklimakan/Taklamakan desert, China, 364, 371
- Taphonomy, 286
- Tephra, 81–3, 106, 271–2, 324–5, 375–7
- Tephrochronology, 81–3
- Termites and soil formation, 268–70, *see also* soils, termites
- Terrace, river, *see* river, terrace
- Thar desert, India, 109
- Thermohaline circulation, 391
- Thermoluminescence (TL) dating, 92–4
- Tibesti Mts., Sahara, 30
- Tibetan plateau
 - glaciation, 379
 - lakes, 203–4
 - uplift, 27, 30, 365–6, 382
- Toba volcano, impact on climate, 81–3, 106, 108–9, 315, 375–7, 383
- Toolmaking, prehistoric, *see* prehistory, stone tools
- Trace gas emissions, *see* greenhouse effect
- Trace element geochemistry, 110–11
- Trace elements
 - in non-marine ostracods, 110, 290
 - in volcanic ash, 376
- Trade Winds, stronger during glacial times, 30, 337
- Transfer functions, 289

- Transgression, *see* lake, transgression
- Travertine, 245, *see also* speleothems
- Tree removal in Australia, effect on salinity, 488
- Tree rings, and drought, 469–70
- Tufa, *see also* speleothems
 - Australia, 253–4
 - Brazil, 255–6
 - definition, 245
 - Somalia, 254–7
 - South Africa, 255–6

- Upwelling, 17
- Uranium-series disequilibrium dating, 91–2, 247

- Vadose zone, 245
- Valley-fill, 195, 274–5, 437–9
- Vegetation cover
 - albedo, Sahelian, 466–8
 - removal, effect of, 174, 466–8
- Vertebrate fossils, *see* fossils
- Volcanism and climatic cooling, 463
- Volcanic dust and drought, 462–4, 472
- Volcanic dust veil index (DVI), 463
- Vostok, 505, *see also* Antarctica

- Wahiba Sands, Oman, 382
- Warm Pool, Indo-Pacific, 427
- Water cycle, *see* hydrological cycle
- Water resources of drylands, 37–40, 52–3
- Water table, *see* groundwater
- Weathering
 - climate, 382
 - deep weathering, 280, 282, 298, 382
 - relict mantle, 280
- West Antarctic ice sheet, 225
 - see also* Antarctica
- West Pacific Warm Pool, 427
- Wetlands, 187–8
- White Nile, 104–5, 173, 178, 184, 195–6, 199–200, 267–8, 351–4
- Wind
 - Gustiness and dust entrainment, 147–8
 - Quaternary fluctuations, 145, 148, 156, 221
 - sand transport, effect on, 117–19
- Wisconsin glaciation, 396
- Würm glacial, 227, 396

- Y chromosome, genetic evidence, 312, 316
- Younger Dryas (YD) stadial, 242, 297, 350–1, 358

